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OUTLINES OF PHYSICAL GEOLOGY

TEXTBOOK OF GEOLOGY

PUBLISHED BY

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PART I

Physical Geology. By the late L. V. Pirsson. Third Edition revised by Wiliam M. Agar, Assistant Professor of Geology; Alan M. Bateman, Professor of Economic Geology; Carl O. Dunbar, Associate Professor of Historical Geology; Richard F. Flint, Assistant Professor of Geology; Adolph Knopf, Professor of Physical Geology; Chester R. Longwell, Professor of Geology. Revision Edited by Chester R. Longwell. Cloth; 6 by 9; 488 pages; 322 figures.

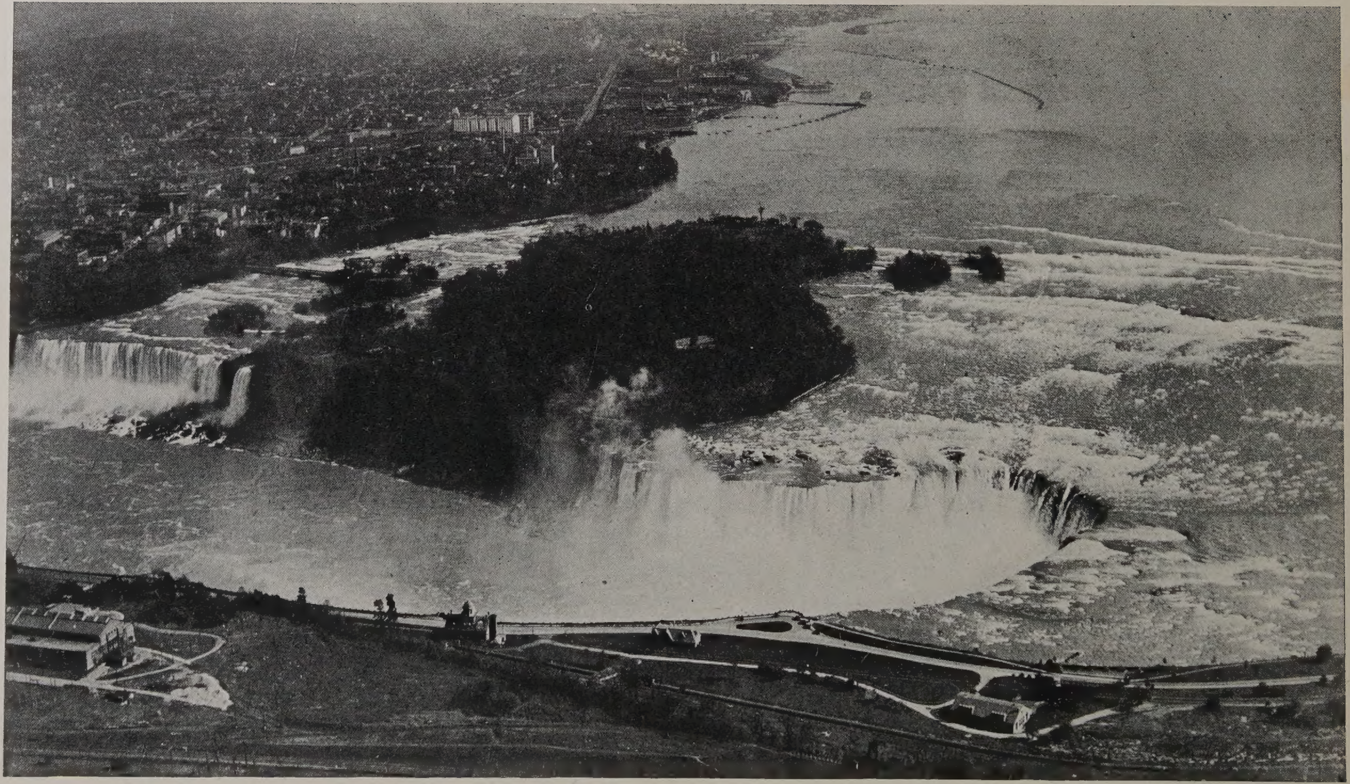
PART II

Historical Geology. By Charles Schuchert. Second Edition, Rewritten and Enlarged. Cloth; 6 by 9; 724 pages; 237 figures in text, 47 plates, and folding colored geological map of North America.

Introductory Geology. By the late L. V. Pirsson and Charles Schuchert. Including "Physical Geology," and an outline of "Historical Geology." Cloth; 6 by 9; 693 pages; 395 figures in text, 26 plates, and folding colored geological map of North America.

Outlines of Physical Geology. Prepared from the Third Edition of Part I of A Textbook of Geology by the late Louis V. Pirsson, and Charles Schuchert. By Chester R. Longwell, Professor of Geology in Yale University. Cloth; 6 by 9; 376 pages; 273 figures.

Outlines of Historical Geology. Part II of Introductory Geology. By the late L. V. Pirsson and Charles Schuchert. Cloth; 6 by 9; 209 pages; 84 figures in text, 26 plates.



Frontispiece

Niagara Falls as seen from the air. View from the Canadian side, looking east. (Courtesy of the Niagara Falls Power Co.)

OUTLINES OF PHYSICAL GEOLOGY

PREPARED FROM THE THIRD EDITION OF PART I
OF A TEXTBOOK OF GEOLOGY BY THE LATE
LOUIS V. PIRSSON, CHARLES SCHUCHERT

BY

CHESTER R. LONGWELL
Professor of Geology in Yale University

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PREFACE

This shorter form of Pirsson's "Physical Geology" is based on the thorough revision made in 1929 by members of the Department of Geology in Yale University. Three chapters of that edition have been omitted here; those entitled "The Interior of the Earth," "Land Forms," and "Ore Deposits." Parts of the material in the first two of these have been incorporated elsewhere in the shorter book. The aim has been to make the treatment briefer and somewhat more elementary. All chapters, with the exception of Chapter I, have been shortened, and in some places the order of topics has been changed, with a view to greater clarity. Errors and ambiguous statements that have been discovered in the 1929 edition have been eliminated. A few diagrams have been redrawn, and new halftones have been introduced wherever it appeared that the illustrations could be improved.

As in the larger text, it is assumed that the student will be given some acquaintance with the common minerals and rocks before he begins the study of weathering and other geologic processes. Logically the text might begin with a chapter discussing these materials of the Earth's crust; but since laboratory materials are required, and since instructors ordinarily use some kind of manual for this study, such an introductory chapter is not considered essential. Methods for studying minerals, with brief descriptions of the commoner species, are given in an Appendix. This short manual does not pretend to give the facts about minerals which the general reader may find most interesting. It is intended as a key whereby the student may learn to recognize and so appreciate the materials with which geology deals. Descriptions of the igneous, sedimentary, and metamorphic rocks appear in the chapters that give systematic discussions of these rock groups. These descriptions can be used in a preliminary laboratory study.

Notwithstanding the extensive changes, most of the material incorporated in the present text was taken directly from the 1929 edition. Disregarding some material added from omitted chapters, the original revision was performed as follows:

Chapters 1, 12, 13, 15 by C. R. Longwell
Chapters 2, 10, 11, 14 by A. Knopf
Chapters 3 and 5 by A. M. Bateman

Chapters 4 and 7 by R. F. Flint
Chapter 6 and Appendix A by W. M. Agar (now of Columbia University)
Chapters 8 and 9 and Appendix B by C. O. Dunbar.

Most of the block diagrams were executed by Phillip B. King.

Chester R. Longwell

New Haven, Conn.,
April 10, 1930.

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OUTLINES OF PHYSICAL GEOLOGY

CHAPTER I

THE SCOPE AND METHOD OF GEOLOGY

Man has a natural curiosity about the Earth, his home. Of what materials is it made? When and how did it come into being, and through what changes has it passed? What has been the story of life on the Earth, and exactly what part has man himself played in the drama? Inquiries such as these have engaged the attention of thinking people from a very early period, as evidenced by the mythologies of the ancients. Some of the old philosophers — Pythagoras, Aristotle, and others — caught brilliant glimpses of the truth; but the science of geology, which strives for the full answer to questions about the Earth and the record of life, had its first consistent development in modern times, with the growth of science in general.

Any understanding of the Earth must begin with some knowledge of the substances that compose it. We are permitted to see only a thin rind of the globe, and therefore all information about the vast interior portion must depend on indirect evidence. However, our most active interest lies in the part that can be explored directly; and this part in itself provides an almost limitless field of study. It is a zone composed of rocks and their constituent minerals. Even to the casual observer it is evident that these materials, exposed in cliffs or in road cuts, tunnels, and other artificial excavations, are highly varied in character; and accordingly it may appear that only a specialist can hope to gain any adequate acquaintance with rocks. Fortunately it has been found that comparatively few types are important quantitatively in the visible part of the Earth, and therefore any educated person may learn without great difficulty to recognize most of the rock masses he may see in the Alps, the Rocky Mountains, or elsewhere in his travels. A study of variations in rock types or of the more detailed features in minerals must of course be left to men specially equipped for the task.

The practical value of recognizing one rock from another is readily apparent from the viewpoint of the professional geologist or the mining engineer. In the search for petroleum and for ore minerals, in the selection of sites for great dams to make storage reservoirs, or in constructing an intricate subway system for a large city, knowledge of rocks and their peculiarities is a vital necessity. But there is a much broader interest in the subject, and the general key to this interest is within easy reach. A visitor to the slopes of Vesuvius or of Mauna Loa sees masses of dark, slaggy rock. It is fairly obvious, even without actually seeing fluid lava, that this dark material was once liquid and flowed down the slopes as a red-hot stream. Continued investigation would convince the traveler that the fire-made or *igneous* rocks are common in many lands and have been formed at many different dates; they constitute an important part of the bedrock beneath us. Again, a brief examination in parts of the high Andes or Himalayas reveals layers of compacted sand or mud which include an abundance of sea-shells. Phenomena of this kind recall the conclusion of Aristotle, " The relation of land to sea changes, and a place does not always remain land or sea throughout all time."

Thus the grouping of rock masses according to the mode and place of their origin is not merely a dull classification for the convenience of scientists; it is the first step in the fascinating game of unraveling the ancient history of a region. In all of its many aspects, the study or practise of geology recognizes this fundamental interest in past events. But just as the proper understanding of human history requires some knowledge of present day social, economic, and political conditions, so the deciphering of events in the geologic past is dependent on acquaintance with processes still operating on and within the Earth. Rocks are not inert monuments to conditions and forces that no longer exist. At active volcanoes we may observe all stages in the development of new rocks from molten material (Fig. 1). In the muds of river deltas and the oozes on sea floors we find the modern equivalents of ancient beds, now greatly distorted and eroded, which furnish the shells and other fossils so common in high mountains and plateaus. Rivers, glaciers, and other surface agencies are etching and slowly wearing away the lands. In some continents the land is being lifted up at a measurable rate, and there is evidence that some mountains are in process of active growth. Thus we are actual witnesses to a constant struggle between Titans; some that work at the surface, striving to tear down the rocky continents, and others within the Earth that persistently oppose the leveling process. Rocks are being destroyed and others are forming to replace them. Activities which can be seen and analyzed have been persistent throughout geologic time, and accordingly we may use the present as a key to the past. All processes now engaged in modifying the Earth are important in the preliminary study of geology.

From the geologist's viewpoint, therefore, the Earth is dynamic and changing; not static and inert. However, since most of the changes are very slow as judged by human standards, it is necessary to form some conception of geologic time in order to appreciate the continuity of Earth-history. The study of geology revolutionizes ordinary notions of time, just as a moderate knowledge of astronomy gives a new vision of space. Our solar system has grand dimensions, and yet in its entirety it is but a point by comparison with the stupendous diameter of the starry galaxy. Similarly we think of the earliest human records as very ancient; but in a geological sense the first appearance of man is a modern event. The oldest relics of primitive men are found only in the superficial soil or other rock débris formed in the latest geologic epochs. In all the time that has elapsed since the oldest civilizations existed, the major landscape features of the Earth have remained essentially unchanged; but we know from the immensely longer geologic record that generations of mountains were made and worn away, before any evidence of man appeared, by the same deliberate forces now at work. The length of time required for such transformations has been almost inconceivably great. How long has it taken for the Colorado River to excavate the Grand Canyon? We are appalled at the realization

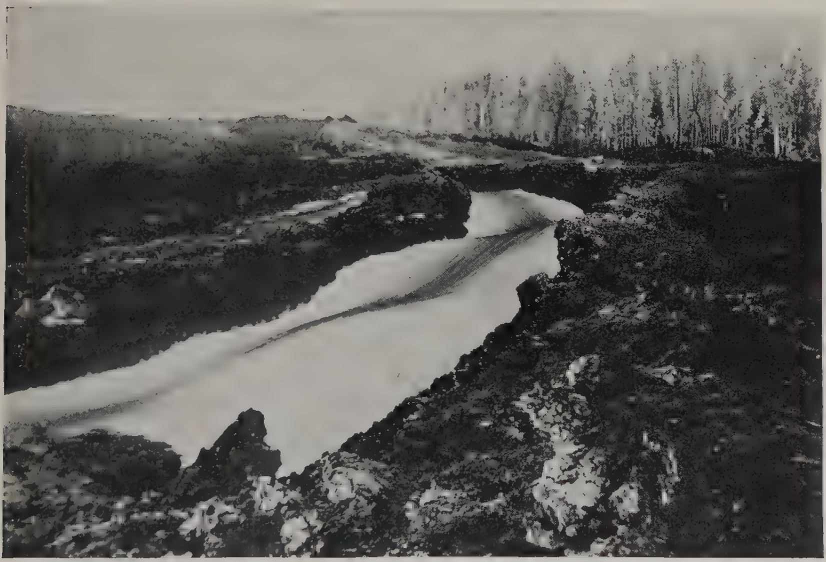


Fig. 1. — Igneous rock in the making. The dark-colored rock is solidified lava; the white band is a stream of fluid lava, flowing toward the observer. (The Alika Flow, 1919. Hawaiian Volcano Observatory.)

that the slow action of water has carried away so many cubic miles of solid rock. Nevertheless the Grand Canyon is a youthful feature. The general record of earlier events is written plainly in the rocks traversed by the trail in climbing the vertical mile from the inner gorge to the outer rim (Fig. 2). Mountains were made and worn down; then the land was submerged beneath the sea for long ages, and continued to sink slowly while limy ooze, mud, and sand built up a deposit thousands of feet thick; this loose material was converted into firm rock; and finally the rising of the wide plateau region high above the sea permitted the cutting of the canyon. Minimum estimates fix the age of the lowest rocks in the gorge at hundreds of millions of years; and yet these rocks may be youthful in comparison with the total age of the Earth. The closest study of the geologic record has revealed "no traces of a beginning, no prospect of an end."

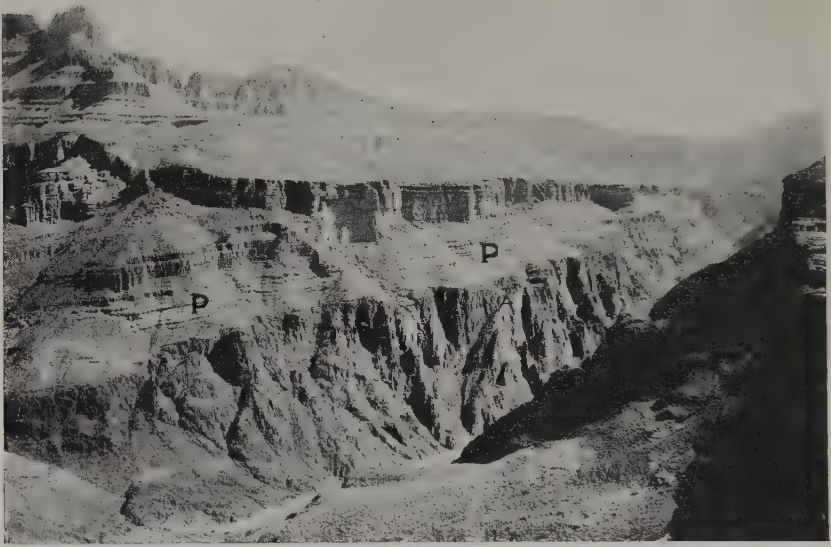


Fig. 2. — Grand Canyon of the Colorado. The sharp inner gorge is cut into extremely ancient crystalline rocks (*A*). The lowest of the sedimentary rocks (*P*) are young by comparison, and yet it is more than 500,000,000 years since these rocks were formed as layers of sand and mud. The cutting of the canyon is a modern event geologically, although it required a long time measured by human standards. (U. S. Geol. Surv.)

This broad, philosophical aspect of geology rests on a secure foundation only because of patient effort by generations of workers in all lands. Prior to the nineteenth century many natural philosophers formed their ideas on geological subjects largely by deductive reasoning, in ignorance of the facts to be learned by observation in the field. Gradually it came to be realized that inductive study, based on all the facts obtainable, is essential for any safe conclusions. A vast array of field evidence has been accumulated. The great mountain ranges have been fruitful fields for geologic investigation, since they furnish the finest exposures of rock formations. But no sources of information are neglected. Geologic features everywhere are examined closely in the field and represented accurately on maps. Mines give opportunity to explore beneath the surface, and deep wells drilled for water or oil yield valuable data. Modern instruments devised to record earthquake waves and to measure the value of gravity have made it possible to draw important conclusions regarding the invisible interior. Slowly but surely the Earth is giving up many of its secrets to the inquisitiveness of man.

Of necessity a field so broad as the study of the whole Earth calls for a division of labor among specialists. Some workers give their principal attention to the rock formations laid down in former seas, lakes, and rivers; some to the volcanic and other igneous rocks; some to the mineral veins and other deposits of economic value. Other subjects of special study are the deformation of rocks by folding and fracturing; the various land forms sculptured by surface agencies; the fossils entombed in rocks; and the minerals that make up rocks of all kinds. All groups of investigators recognize unsolved problems, and by their united efforts continue to widen the frontiers of the science as a whole. Like any other growing subject, geology extends beyond the lighted zone of proved fact. It has also a large twilight zone of inference and probability into which the full light of investigation continues to spread slowly; and beyond this a region of shadow and complete darkness, relieved only by scattered flashes of speculation. The speculative side of the subject is in some respects the most fascinating, but it may also be dangerous to the uninitiated. A comprehensive discussion of geology takes account of fact, inference, and speculation, but distinguishes carefully between them. Careless broadcasting of hypotheses as if they were proved facts has given rise to numerous popular misconceptions about the Earth and its life.

The study of geology begins logically with an introduction to the important kinds of rocks and minerals. It proceeds to an examination of the forces that act on the outer part of the Earth, and the changes produced by these dynamic agencies. Finally, the keys provided by this preliminary study are used to explore the long geologic record contained in the rocks.

CHAPTER II

A GENERAL VIEW OF THE EARTH

The Earth and Its Neighbors. — In a comprehensive study it is well to get a detached view of the Earth in its astronomical setting. The Earth is one of a group of planets that revolve around a common central orb — the sun. Some of these, like Jupiter, are much larger than the Earth; some like the asteroids, or minor planets, are much smaller; some are much nearer the sun, others farther away. The group has nearly a common plane of revolution about the sun, as suggested in Fig. 3, and this fact is thought to have an important bearing on the origin of the solar system. The Earth and other planets were born of the sun; according to the modern theory of cosmogony, which we owe to the American geologist, Chamberlin, the substance of the planets was expelled from the sun under the influence of the disruptive pull of a passing star several times larger than itself; but how this ejected matter was aggregated to form the planets is a problem on which ideas are still far apart. The common plane of revolution of the sun's satellites is possibly an inheritance from the biparental origin of the system. The birth time of the Earth was probably at least 2000 million years ago, and since that time the Earth has been gradually acquiring its present character. A vast span of time has thus elapsed, and that portion the record of which is written in the rocks is referred to as "geologic time."

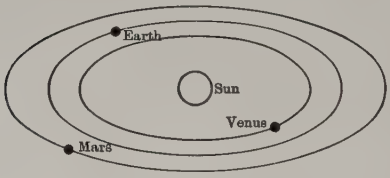


Fig. 3. — Planets revolve about the sun in nearly one plane, as suggested by the orbits of three of them.

The sun is called by astronomers a G-type dwarf star; by this they mean that it is a comparatively small member in an important group of yellow stars, distinguished from others that emit red, white, or bluish light. Its surface temperature is approximately 6000° C. as determined observationally; the temperature at its center is estimated to be 40,000,000°. The geologic record shows that the sun has been supplying light and heat to the Earth at an approximately uniform rate during hundreds of millions of years. Up to the early part of the present century the other sciences could not adequately explain how the sun had been able to maintain this prodigal expenditure of energy. It is now considered probable that its fires are stoked either by utilizing subatomic energy or by the conversion of matter into radiant energy — the annihilation of matter. According to Eddington, the transmutation of the elements would keep the sun going 10,000 million years, but the annihilation of matter would give " a very ample time scale."

The mass of the sun is 332,000 times that of the Earth. Even counting in the masses of the other planets, practically all of the mass of the solar system is in the sun. Therefore the sun is the overwhelmingly dominant member of the system. Carrying with it the Earth and the other planets, the sun is travelling through space at the rate of 12 miles a second.

The path of the Earth about the sun is not a circle, but an ellipse, one of whose foci is the sun. The deviation of the ellipse from a circle, however, is relatively small; the average distance of the Earth from the sun is nearly 93 million miles. In consequence of its elliptic orbit, the Earth is 3 million miles nearer the sun on January 1 than it is on July 1. This fact causes the summers to be somewhat cooler and the winters somewhat warmer in the northern hemisphere than they would otherwise be. At the present time summer, as measured from the vernal equinox on March 21 to the autumnal equinox on September 23, is 7 days longer than winter. In the southern hemisphere, however, winter is the longer season. Owing to the precession of the equinoxes, this condition will be reversed between the two hemispheres 10,500 years hence. Because of the perturbing effect of the other planets, the eccentricity of the Earth's orbit varies during a period of 500,000 years, and at maximum eccentricity the Earth is 13 million miles nearer the sun in summer than in winter, thereby causing short hot summers and long cold winters. At that time the winters will be 35 days longer than summer. As we shall see later, these astronomic conditions have been held by some to be sufficient to produce great climatic changes, to determine cycles of sedimentation, and even to bring about the several glacial epochs that have affected our planet during its long career.

Besides revolving around the sun, the Earth is spinning on its polar axis, each rotation in 24 hours giving rise to day and night. The axis of rotation is not perpendicular to the plane of the Earth's orbit but inclined to it at an angle of $66\frac{1}{2}°$, and this inclination gives rise to the seasons, summer and winter, alternately in the northern and southern hemispheres according as the axis is pointed toward the sun or away from it.

The Earth, as we have seen, is a very insignificant fraction of the solar system, and for that matter the solar system itself is but an insignificant fraction of the universe. The sun is a modest star in a stellar system whose members are numbered in thousands of millions. At immense distances beyond this system — our " universe," as it is currently called in astronomy — are other gigantic systems. According to Sir J. H. Jeans, " the farthest astronomical objects whose distances are known are so remote that their light takes over 100 million years to reach us." Nevertheless, throughout this vast extent, with its island-universes containing myriads of stars in various stages of development, the same general physical laws that we know on the Earth appear to govern. Gravity operates in the same manner; light is transmitted everywhere by the same kind of vibrations; the spectroscope tells us that the same chemical elements occur in distant stars as on our Earth. Moreover, the meteorites that our planet gathers in its journey through space and which appear to be the disrupted fragments of former worlds are made of substances identical with those found on the Earth. Consequently there appears to be a unity of law and a uniformity of material throughout space, and we feel justified in assuming that facts and reasoning derived by astronomical study of the other heavenly bodies may also be applied in our study of the Earth.

Size, Shape, and Composition of the Earth. — The science of *geodesy,* which is concerned with exact measurement and mapping of the Earth's surface, has determined with precision the dimensions and the form of the globe. The Earth is not a true sphere but a spheroid flattened at the poles, so that the axis on which it rotates is slightly shorter than the equatorial diameter. The polar flattening (oblateness) of the Earth and the other planets is fully explained by the principles of celestial mechanics. It depends on the centrifugal force due to rotation and on the internal constitution of the planet. By internal constitution is meant the composition, density, and distribution of the density within the planet.

By precise physical experiments we determine the " constant of gravitation." The method involves essentially the measuring of the force with which the Earth attracts a body of known mass. The result gives the total weight of the Earth; and since its size also is known, it is a simple matter to compute the average density. This value, arrived at by many experimenters, is 5.52; that is, an average sample of the Earth weighs five and a half times as much as an equal volume of water.

Direct determinations of density, using all kinds of rocks known at the surface, give an average value of 2.7. As this is less than half the density of the whole Earth, it appears that the interior must consist of much heavier material than the outer part. Indirect evidence suggests a central core of metal, with a diameter about half that of the entire

Earth; and outside this core a thick zone of rocks much heavier than those that are common at the surface.

Other Facts About the Earth. — Volcanoes and hot springs indicate high temperatures at depth locally, and direct measurements in deep mines and wells suggest a universal increase in temperature downward. The rate of this increase varies between wide limits. In some places it is as high as 1° F. in 30 feet; in other places, as in the deep gold mines of the Transvaal, it is only 1° F. in 250 feet. The average for all observations is about 1° F. for 60 feet. Therefore mines that reach down several thousand feet are uncomfortably warm.

The determination in recent years that all known rocks contain appreciable quantities of radioactive material is of great importance. It appears certain that the distribution of these materials must be confined to a comparatively shallow outer zone. Radioactive elements such as uranium and thorium break down at a constant rate, producing heat; and as rocks conduct heat away at an extremely slow rate, it can be calculated that the existence of these elements even at moderate depths would result in permanent fusion. It is known, however, from study of earthquake waves and from other evidence, that the Earth is solid and rigid, at least to a depth of nearly 2000 miles. Granites contain higher percentages of radioactive substance than other known rocks, and hence it is argued that granites, which are the predominant rocks in the continents, are limited to shallow depth.

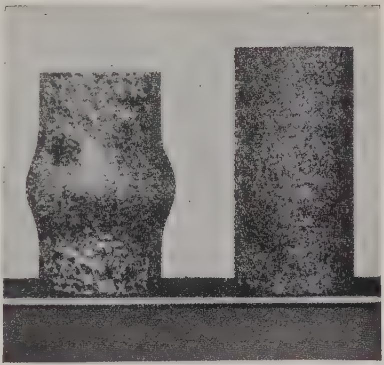


Fig. 4. — Rock flowage in laboratory test by Adams and Bancroft. The rock column shown on the right was placed in a strong steel jacket and subjected to great pressure (up to more than 100,000 pounds per square inch) by two pistons fitted to the ends of the column. After the experiment the rock cylinder had the form shown at the left.

Behavior of Rocks Under Pressure. — Since ordinary rocks are brittle, they will fracture and crush under high pressures in the laboratory unless special precautions are taken. When a rock specimen is confined on all sides, and subjected to very intense compression, its size decreases slightly; in other words, rocks are somewhat compressible. If a marble core is fitted into a cylindrical opening in a strong steel jacket, and subjected to enormous pressure by means of steel pistons, in time the walls of the steel jacket are bulged outward. By cutting away the jacket it is found that the marble core has been shortened and thickened, *but not crushed* (Fig. 4). Evidently it has *flowed* slowly, as if it were a plastic substance. Since rocks deep in the Earth are confined under high pressure, it is argued that under certain conditions these rocks are deformed by plastic flow. Strength and rigidity are only relative terms.

Relief Form of the Earth

General Features. — The irregularities of the Earth's surface, or its relief, divide naturally into major and minor groups. The major relief features are the continents and the ocean basins. The continents are regarded as being partly submerged beneath the oceans, and the submerged portions are known as the continental shelves. The minor relief features of the continents are the mountains and plateaus as opposed to valleys and basins; and the minor relief features of the ocean basins are islands and submarine ridges as contrasted with the profound troughs or *deeps* in the ocean floor.

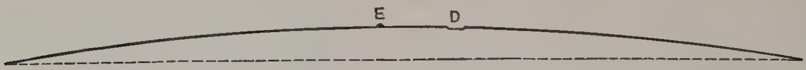


Fig. 5. — Mt. Everest in comparison to the size of the globe. Part of the arc of a globe with radius of one foot is shown; on this Mt. Everest, *E* ,would be about $\frac{1}{50}$ of an inch high. *D*, in a similar way, shows the greatest depth of the ocean.

The average height of all the lands above sea-level is 2300 feet. North America averages about 2000 feet. The average depth of the oceans is about 13,000 feet. The highest elevation of the land, Mt. Everest in the Himalayas, is 29,000 feet; the greatest known depth in the ocean, in the Pacific, is 35,400 feet. Thus the greatest difference in relief exceeds 64,000 feet, or more than 12 miles; and the average height of continents above the ocean floors is nearly 3 miles. Relative to its size, however, the globe has an extremely small relief, and it is therefore comparatively smooth; Fig. 5 shows its greatest roughness.

The relief features of the land are the *plains*, such as the Atlantic Coastal Plain; *plateaus*, such as that of the Colorado; and *mountains*, like the Appalachians extending from Canada to Alabama. In regard to the grouping of the relief forms of the Earth, certain facts are of interest and importance. The continents as a rule consist of basins that are bordered by mountain chains near the coastal rims, whereas the ocean basins generally have the reverse arrangement, the deeps being near the continents and the submarine ridges, or upswells of the bottom, being in mid-ocean. Some of the highest and most important ranges on the edges of the continents border the greatest deeps in the ocean floor, as for instance the Andes in South America and the partly submerged mountain chain that forms the Japanese islands and is the real eastern border of the continent of Asia; close to these ranges the ocean floor descends to great depths. It is not meant to imply, however, that mountains occur only at the continental edges, for some of them extend in a wide zone far into the interior, as in western North America, or form systems crossing a continental mass, as in Asia.

Relation Between Density and Form. — Theoretically the surface of the Earth would be quite smooth if the material below the surface were uniform in its character. In reality, we know that the rocks exposed to observation differ considerably in composition and in density. By geological investigation and by highly technical instrumental determinations, a strong probability has been established that the rocks underlying the oceans are appreciably denser than those composing the continents. Moreover it is concluded from careful geodetic study that the great mountain ranges of the Earth are composed of or underlain by material that is slightly deficient in density compared with the crust as a whole. It would appear from these facts that the larger surface irregularities are not haphazard, but have a fundamental cause; that differences in surface elevation are compensated by differences in rock density. In other words, the lighter parts of the crust project above the heavier parts, just as icebergs rise above the water around them. This condition of balance in the crust is called *isostasy.*

The differences in density that make possible the larger irregularities of the Earth's surface are of fundamental importance. If the outer part of the Earth were uniform in composition there could be no land life, because the oceans would cover the entire surface.

The Outer Zone of the Earth; Rocks

It is chiefly the outer zone of the Earth about which we have extensive and positive information. This outer zone is known as the crust; it is built of rocks, which are discontinuously covered by a thin mantle of soil. Because " crust " is thought by some to connote that the Earth has a liquid interior, the term lithosphere has been coined to avoid this implication, but we shall use the simpler term preferably. It is upon the crust that we live and exert our activities; we penetrate into it for fuels of various kinds, for metals, water, building material, and other mineral resources, all of which are essential to the physical side of modern civilization. A thorough knowledge of the component parts of the Earth's crust and its structure is consequently of the highest importance. The component parts of the crust are rocks, and we shall begin our inquiry by a study of the different kinds of rocks and the varied modes in which they occur.

Definition and Classification of Rocks. — The word rock, in the language of geology, means the material that composes one of the individual parts of the Earth's outer shell. According to their mode of origin, rocks are divided into three main groups: the *igneous* rocks, made by the solidification of molten material (Fig. 1); the *sedimentary*, or bedded rocks, formed from sediments that were deposited chiefly by water (and to some extent by air and ice) (Fig. 141); and the *metamorphic* rocks, formed by certain processes acting within the Earth's crust on preëxisting rocks and partly or wholly destroying their original characters and producing new ones, so that the resultant rocks are best considered as constituting a separate group (Fig. 244).

Thus we have three groups:

I. Igneous Rocks — consolidated molten masses.
II. Sedimentary Rocks — formed from sediments deposited by water, air, or ice.
III. Metamorphic Rocks — secondary, derived from preëxisting rocks.

Three-fourths of the land area of the globe is underlain directly by sedimentary rocks and the other fourth by igneous and metamorphic rocks. Although the sedimentary rocks thus preponderate in the visible part of the crust, they are essentially a veneer, a mile or less thick on the average. The foundation rock of the continents is largely igneous rock (granite).

Some acquaintance with the important kinds of rocks and component minerals, from study of typical specimens, is required for an adequate understanding of the discussion in succeeding chapters.

CHAPTER III

THE ATMOSPHERE; WEATHERING AND SOILS

General Functions of the Atmosphere. — The atmosphere is directly essential to life on the Earth, and in addition it enables the sun and the rain to bring about life-giving changes on the Earth's surface. Without the atmosphere the Earth would be boiling hot by day or freezing cold by night, and devoid of life, as the moon is supposed to be. There would be neither wind nor rain nor bodies of fresh water. The lands would be rugged and desolate. Winding streams and soft landscapes mantled with colored soils and vegetation, and other vistas pleasing to the eye, would be lacking. For all these things, as we shall see, result directly from the presence of the atmosphere. Due to the rotation of the Earth and unequal heating by the sun, movements are set up in the atmosphere that account in part for our climate and weather. The wind itself builds up structures such as sand dunes, or abrades the rocky surface by the sand it carries along. But far more important are the chemical action of air and moisture, and the mechanical effects of changes in temperature, or frost action, aided by plant and animal life, that cause rocks to change into soil. This combination of processes, by which rocks at the Earth's surface break to pieces and decay, is called *weathering*.

Character of the Atmosphere. — The atmosphere consists of three chief constituents in the following quantities by weight: — nitrogen, about 75 per cent; oxygen, about 23 per cent; and argon, about 1.4 per cent. There are also present in small quantities the inert gases krypton, xenon, helium, and neon, but these are of no importance geologically. In addition, water vapor, carbon dioxide, hydrogen dioxide, ozone, ammonia, and traces of hydrogen, sulphur, organic matter, and suspended solids, are present in varying quantities. But of all these constituents only oxygen, water vapor, and carbon dioxide play an outstanding part in geologic processes.

The oxygen, in addition to its obvious function of sustaining life, also enters into chemical changes that are important geologically. Wherever animals breathe or fire burns, oxygen is withdrawn from the air and locked up in compounds such as carbon dioxide. Also the weathering of rocks involves the production of iron oxides such as limonite, in which oxygen is absorbed from the air. But growing plants, on the other hand, liberate some oxygen from carbon dioxide.

The carbon dioxide content in the atmosphere is relatively constant at 3 parts in 10,000 by volume. Though the quantity is small, its geological importance is great. Carbon dioxide is being added to the atmosphere from volcanic and mineral spring emanations, the combustion of fuels, the respiration of animals, and the decay of organic matter. It has been estimated that the consumption of coal annually returns to the atmosphere about 1/1000 of its present content, so that in 1000 years the amount of carbon dioxide in the atmosphere would be doubled from this source alone. But other operations are as continually diminishing the carbon dioxide content, notably its extraction from the air by growing plants, and its great consumption in the weathering of rocks.

The carbon dioxide and water vapor of the atmosphere also exert a beneficial effect on the Earth by causing part of the solar heat to be retained. Thus they help make the Earth hospitable to living things, and they aid chemical reactions that are important in the formation and the destruction of rocks.

CLIMATE AND WEATHER

It is obvious that rock weathering will proceed differently in regions where rain falls intermittently throughout the year and in places where rainfall is seasonal; also the effect of these processes will be unlike in an arid climate and a humid climate. Consequently, a knowledge of climate and weather is essential to a proper understanding of geologic processes that produce far-reaching changes on the Earth's surface.

Movements of the Atmosphere. — Air heated in the equatorial regions expands, rises, flows poleward, and settles down some distance north and south of the equator, and then moves over the surface toward the equator. The colder polar air tends to move equatorward. This circulation resembles that of the hot water in a house-heating system, and like the latter is induced and maintained by temperature differences. A great atmospheric circulation is thus set up. If the Earth did not rotate, and were smooth, and if its temperature varied uniformly from equator to poles, just such a simple atmospheric circulation would prevail. But this ideal simplicity does not exist. The rotation of the Earth from west to east throws this otherwise simple circulation very much askew, so that the air moving toward the poles is deflected eastward and that moving toward the equator is deflected westward. There are thus set up in the lower part of the atmosphere belts of *prevailing* or *planetary winds*. Those in the equatorial regions, roughly between latitude 30° N. and S., are the *easterlies* or familiar *trade winds*, and those in the middle latitudes are the *westerlies*.

But in the middle latitudes there are, in addition, minor or secondary movements known as *cyclonic storms,* which affect animal and plant life from day to day and give rise to our weather. The westerlies of these latitudes are not simple prevailing winds always from the west. Here the equatorial and polar currents intermingle and, due in part to imperfectly understood effects of the upper atmosphere, and in part to unequal heating and cooling, circular air movements called *cyclones* (low pressure areas) and *anticyclones* (high pressure areas) are set up. These are essentially thin, flat discs of moving air from 500 to 1000 miles across, that sweep easterly at an average velocity of 500 to 800 miles a day. Within these discs, the movement may be pictured as whirlwinds in which, in each of the low pressure areas (the *Lows*), a rising spiral of warm air rushes anticlockwise toward the center in response to the low pressure; whereas in each of the high pressure areas (the *Highs*) a descending column of cold air spirals toward the outside in a clockwise direction, being pushed out, as it were, by the high pressure. In the southern hemisphere the directions are reversed. About three Highs and three Lows may be expected to pass from west to east over any point in central and northeastern United States every two weeks. The direction of the wind at a given place at any particular time will depend upon which part of a Low or a High covers it. Thus the weather in the middle latitudes is continuously changing.

Atmospheric Moisture. — Water in the form of vapor is always present in the atmosphere, though its amount varies greatly from time to time and from place to place. When the atmosphere carries its full capacity it is said to be *saturated,* but this condition depends largely upon temperature, since warm air can contain more moisture than cold air. If air is only one-half saturated for a given temperature, it is said to have a *relative humidity* of 50. The average relative humidity over the oceans is about 85; over the lands it is considerably less. If air with a relative humidity of, say, 60 is cooled steadily, though there is no change in the actual amount of water present, the relative humidity will increase up to 100, or saturation, and further cooling will cause *condensation.* Upon this simple principle depends the control of precipitation, whether in the form of dew, fog, rain, or snow.

In nature there are several possible causes for cooling of moisture-laden air, such as radiation, moving of cooler air to it, or rising of the moist air into higher altitudes. The latter cause, with the effects of attendant expansion, is the commonest. No surprise can be felt, therefore, at the heavy rainfall on mountains that lie in the path of air currents moving from the sea, for we know that high ground forces the air to move upward, where cooling by expansion aids condensation into clouds and rain or snow. The same thing occurs when warm, moisture-laden air moves upward in the center of a low-pressure area, and that is why cloudy weather or precipitation accompanies Lows.

Conversely, if the air is warmed, its capacity to absorb and hold moisture is increased and it becomes undersaturated. Consequently, the cold air that falls in the center of a High becomes warmed by compression and is a drying wind from which no precipitation occurs. Similarly, air that falls on the lee side of a mountain range becomes warmed in its descent; if the fall be rapid and great, the skies are clear, the air is warmed, and any surface water in its path is rapidly absorbed.

Atmospheric Temperatures. — The atmosphere receives most of its heat from the sun and, as pointed out above, this solar heat is retained largely through the action of carbon dioxide and moisture in the air. More heat is absorbed from the direct rays of the sun near the equator than from the slanting rays near the poles, and consequently the temperature, generally speaking, decreases from the equator to the poles. But the uniformity of this arrangement is disturbed by ocean currents and the disposition of land and sea. Land absorbs more of the sun's heat than does water, and gives it off more readily; consequently it becomes warmer by day and cooler by night than does water. Water retains more of the heat it receives and tends to store it up until it is in part distributed by currents. For these reasons land climates are more variable than oceanic climates, and maritime regions are milder in winter and cooler in summer than continental areas. In the middle latitudes, local changes in temperature are sudden and pronounced, owing to the alternation of Highs and Lows.

Weather. — In middle latitudes the procession of Highs and Lows brings changing winds and temperature, and clear or rainy spells. Thus the weather moves and varies. In North America a low pressure area moving generally eastward may in its broad reach cover a considerable part of the eastern United States and southern Canada (Fig. 6). Winds from the south and east, after their sweep over warm sea water, deliver moisture to a Low. When they ascend in the cyclonic center, cooling takes place and condensation soon occurs, so that cloudy or rainy weather accompanies the Lows. This eastward movement of Lows, then, gives a generous rainfall to the eastern part of the continent. Occasionally a Low remains nearly stationary, and then rains may be so continuous that floods result.

As a Low passes and a High follows, the warm winds from the southeast and south shift to the west and northwest and the temperature falls rapidly. This causes the cool spells of summer and cold waves of winter. The descending cool air is undersaturated; drying winds and clear skies result.

Climate. — If the Earth had a smooth, homogeneous surface and no atmosphere, there would be simple solar climates whose distribution would correspond with latitudes. While it is true that latitude is the most important single factor controlling climate in so far as temperature is concerned, there are other factors that profoundly modify it. For example, compare the equable climate of the British Isles, tempered by winds from the warm Gulf Stream, with that of Labrador at the same latitude, washed by the chilly Labrador current. A region in which the winds blow from an adjacent ocean has, for its latitude, mild winters and cool summers and small daily variation in temperature.

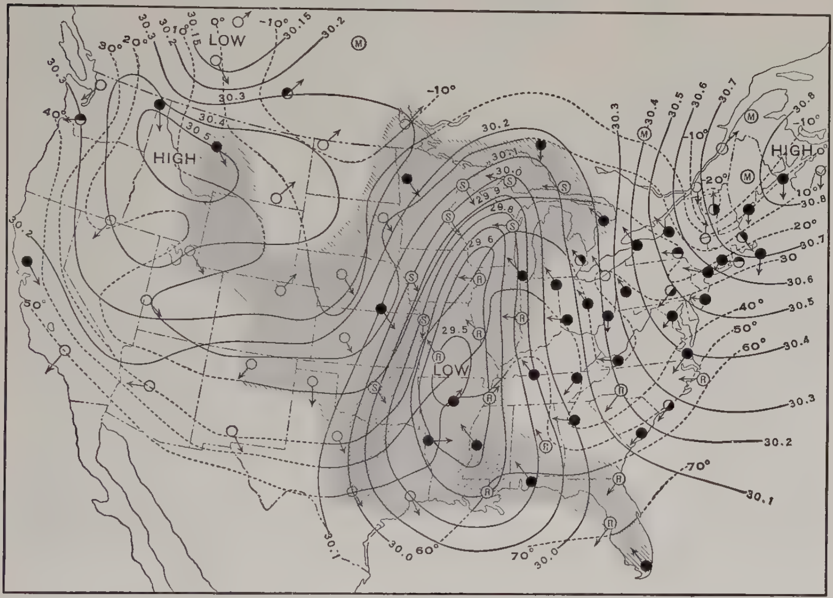


Fig. 6. — Weather map of the United States for Jan. 5, 1929. Shows high and low pressure areas, and a storm in the Mississippi Valley accompanying the "Low."

Continental climates are marked by great variation in temperature, rainfall, and barometric pressure. In the United States, for example, a thin maritime belt (Fig. 7) fringes the northwest coast where extremes in temperature are few and rainfall is plentiful. The prevailing westerly winds carrying moisture from the Pacific Ocean are forced high by the coastal ranges, whose western slopes are thus well watered. Where the dry winds drop suddenly on the eastern lee slopes, there are arid and semiarid belts where temperature changes from night to day are pronounced. The contrast with the coastal belt is striking. Throughout the Great Basin region arid or semiarid climates prevail and the daily and annual variation in temperature is extreme. The climate of the Mississippi drainage basin and the Eastern States is affected strongly by cyclonic storms and proximity to the Atlantic Ocean. Sudden changes in temperature are frequent, and clear and rainy weather alternate throughout the year. The New England States, over which most of the cyclonic storm paths converge, have particularly changeable weather, with fairly evenly distributed precipitation throughout the year, and pronounced temperature changes. Warm and cold spells alternate with surprising suddenness, even in midwinter, due to a High following a Low. This is typical of a continental climate in middle latitudes. An extreme example of the continental climate is central Asia, which is not only remote from the oceans, but is also cut off from ocean winds by mountain barriers.

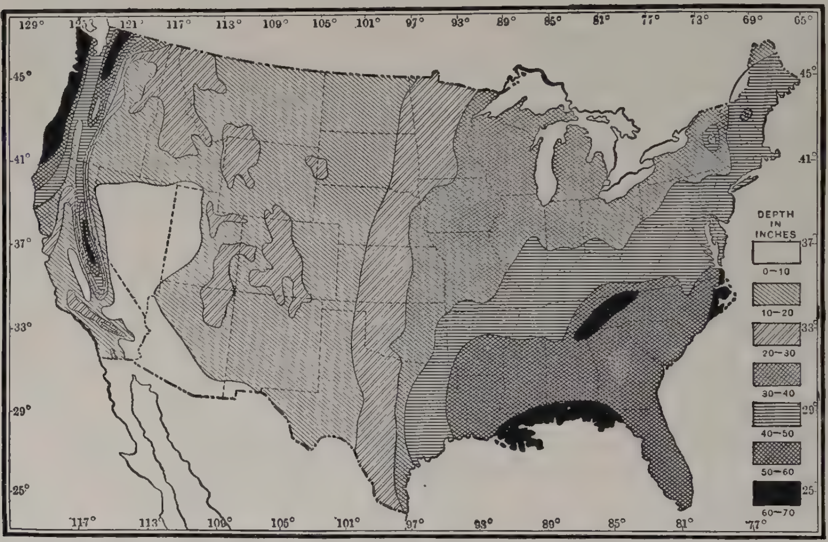


Fig. 7. — Rainfall map of the United States. (U. S. Geol. Surv.)

Factors that Change Climates. — This brief discussion makes no pretense to be an adequate treatise on the complex subjects of climate and weather. The intention is to emphasize the factors that control climates, and to make it clear that if the controls be changed, then a different climate for a given place will result. Examples are changes in the distribution of land and water, with a shifting of the ocean currents;

the formation of a new mountain range on the land; changes in solar radiation, or other astronomic changes; and changes in the carbon dioxide content of the atmosphere, or in the atmospheric circulation. The studies in geology to follow will teach us that in the past just such changes of climatic controls have taken place.

Climates of the Past. — The records of geology show that there have been profound changes of climate in the course of the Earth's history, and such changes are probably still in progress. Fruitful lands of today were once barren deserts; subtropical plants thrived in what are now cold northern regions. In the period just preceding our own, widespread glacial conditions prevailed over the regions of high latitudes, and ice sheets spread over lands that now enjoy temperate climates. Still earlier, glaciation spread over what now are subtropical parts of Africa (Fig. 8), Australia and India.

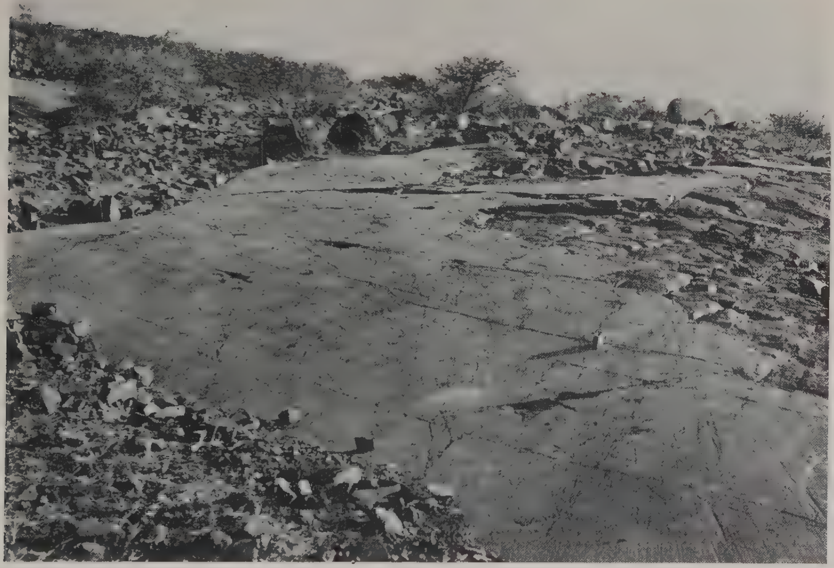


Fig. 8. — Proof of ancient glaciation near Kimberley, South Africa. The rock floor was grooved and polished by the movement of the great ice sheet. Boulders and pebbles lying on the surface were the tools with which the scratches were made. This region now has a subtropical climate. (Longwell.)

These climates of the past, in their changing character throughout the ages, have influenced notably the development of life upon the Earth, and have affected rock weathering and other geologic processes.

WEATHERING

The attack of the atmosphere upon the rocks is twofold — mechanical and chemical. The first produces *disintegration* of the rocks by mechanical agencies such as frost action. The chemical attack results in *decomposition* of the rocks, and a certain amount of this is essential to form good soil. Both kinds of destruction usually operate together and each is aided somewhat by the work of plants and animals. The dominance of one or the other depends to a large degree upon the nature of the climatic provinces in which they operate. Together they constitute a rather complex set of processes called *weathering.*

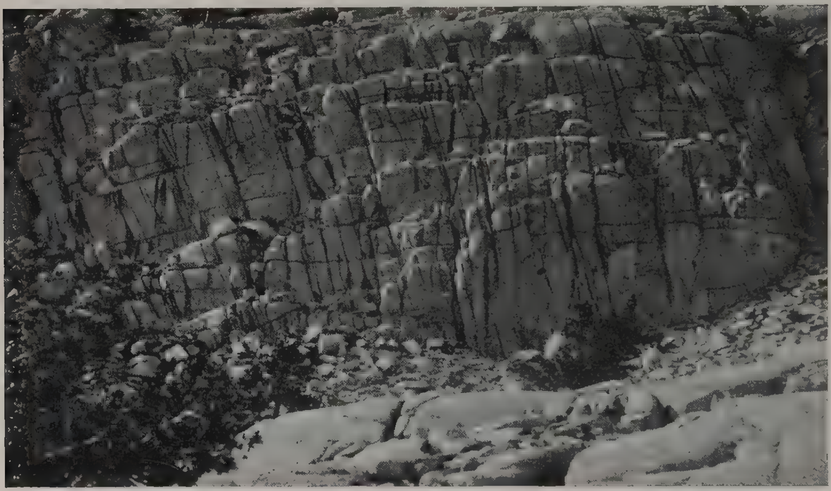


Fig. 9. — Shattered nature of the bedrock which facilitates disintegration and decomposition by allowing ready entrance of water, air, and plant roots. (U. S. Geol. Surv.)

Weathering is aided greatly by the many fissures that penetrate the bedrock, (Fig. 9). Even the mineral grains of the rocks are more or less filled with cracks. These openings are potent factors in promoting disintegration and decomposition, for they allow ready entrance of air and water and plant roots into the rock.

Mechanical Agencies

The Expansive Force of Freezing Water. — Water pipes that burst upon freezing are a familiar manifestation of the expansive force of freezing water — a force that exceeds 2000 pounds per square inch. In the days of muzzle-loading cannon, captured enemy guns were frequently disposed of by being filled with water and left to freeze, by which means they were split from end to end. The effect upon rocks is similar; water trickles into the numerous crevices, joints, and pores (Fig. 9), and upon freezing, pries off small pieces or large blocks. The latter are in turn broken into smaller pieces. In the same manner porous sandstones may be completely reduced to sand. This action goes on whenever water freezes in a confined space. Naturally it is most pronounced where freezing and thawing occur frequently, as in most temperate regions in the fall and spring. This is particularly the case on high mountains where, almost regardless of latitude, thawing by day and freezing by night goes on over a considerable part of the year. Anyone who has camped on a steep mountain on a frosty evening may have heard the startling noise of falling rocks dislodged by this process. Débris formed in this way contributes to the formation of " slide rock " (Fig. 14) on the mountain slopes. Those who climb mountains are familiar also with the masses of broken rock on the tops of ridges. These are impressive illustrations of the effectiveness of frost action on larger topographic features. The results of this process are also clearly evident on many buildings and monuments throughout the northern States.

Frost action, even by itself, is an important factor in mountains and in temperate and cold climates, in helping wear away projecting rock masses. Further, it provides smaller fragments upon which chemical attack may be made more effectively. But it must not be assumed that all heaps of broken " slide rock " are caused only by frost action, for several other processes of rock disintegration produce similar results.

Changes in Temperature. — Most substances when heated expand, and if they are good conductors of heat, like metals, the expansion extends well beyond the place of heat application. An iron rod cannot be held with the free end in the fire without discomfort to the hand; but a piece of rock can be so held, because rock is a poor conductor of heat. Because of this fact the heated rock expands unequally in its different parts, and is likely to be disrupted by the strains that result. The heat of a forest fire causes large flakes and spalls to break from granite and other exposed rock surfaces. Polished faces of stones in buildings are damaged by fire in the same way.

Ordinarily the changes in temperature experienced at the Earth's surface are small. Heating by the sun is most extreme in desert regions, where rocks grow uncomfortable to the touch during a summer day. Since rapid cooling at night is also characteristic of arid climates, the daily range in rock temperatures is comparatively large in a desert.

Exceptionally it is as much as 100° F. There is a common belief that such repeated heating and cooling is the chief cause of rock disintegration in arid countries; but this assumption has not been proved. Laboratory experiments using even a greater temperature range than that recorded in any desert have not caused the slightest breaking of rock samples. Possibly the daily repetition of small temperature changes during thousands of years finally causes rocks to weaken, with the results of disintegration observed in deserts. It is probable, however, that slight chemical effects, to be explained later, are at least partly responsible for the breaking up of rocks in arid regions.

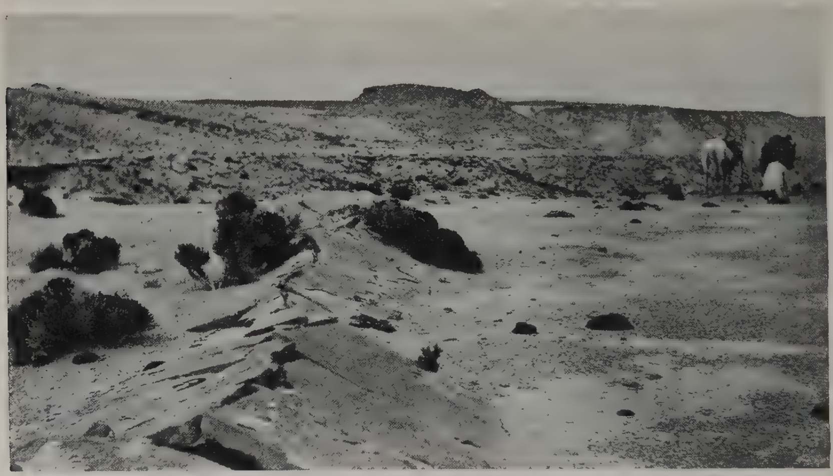


Fig. 10. — Buckling in sandstone layers due to expansion from heating by the sun. Wyoming. (U. S. Geol. Surv.)

Heating of sedimentary rocks causes local buckling and rupturing of layers (Fig. 10). This effect is especially noticeable in the floors of quarries where strata have been exposed recently to direct heating by the sun.

Plants and Animals. — The wedging apart of rocks by the roots and trunks of growing trees is an important process (Fig. 11). Likewise, the rootlets of plants and shrubs insinuate themselves into little crevices of bedrock and boulders, and slowly enlarge the openings. In the course of time the amount of work done in this way must be great, but its chief importance is in helping chemical attack to proceed more readily.

Animals such as moles, gophers, worms, and ants, by making holes and burrows in the soil, aid disintegration to a slight degree. Their work enables the other agencies to come more readily into contact with covered rock surfaces. The most destructive animal, however, is Man. He has felled the forests and otherwise destroyed the protective covering of plant life over the soil, allowing it to be carried away by rain or wind and thus expose fresh surfaces of rock to processes of weathering.

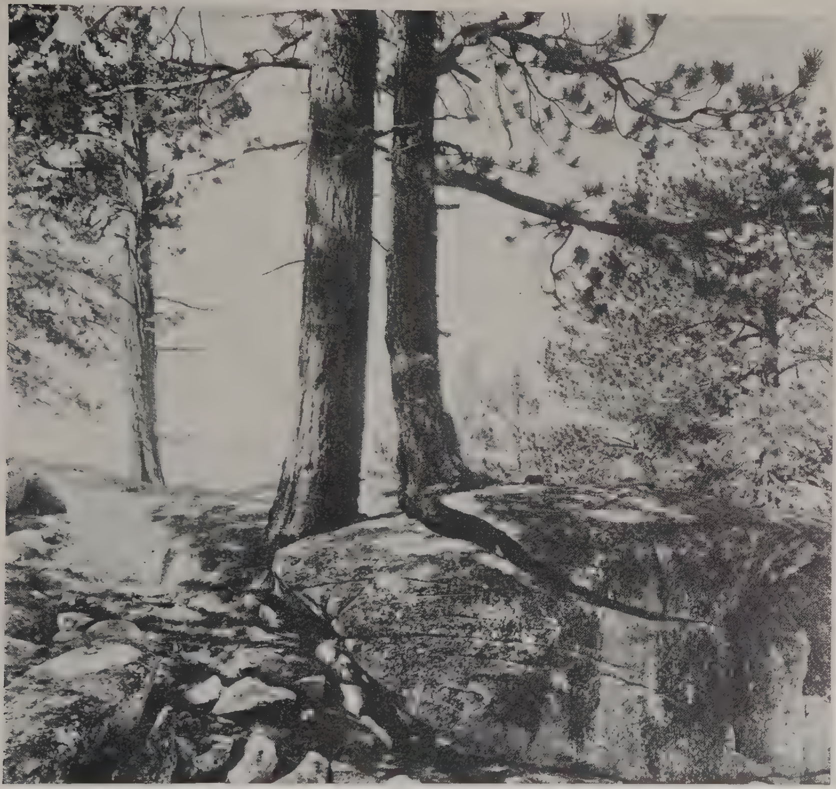


Fig. 11. — Splitting of rocks by the growth of trees. (U. S. Geol. Surv.)

Chemical Agencies

Chemical agencies of weathering, like mechanical agencies, are essentially superficial in their work. They act only on a thin skin of the outer rocky crust, and lose their effectiveness at comparatively shallow depths. Their work is inextricably interwoven with that of the mechanical agencies, and it is only for convenience of discussion that we consider them separately. The result of their combined attack is soil, which is the basis of human life.

Decomposition or decay may act directly upon solid bedrock, but it proceeds much more readily upon rocks previously fragmented by disintegration processes. It goes on more slowly than disintegration, and is favored by moisture and heat, and retarded by cold. Consequently, unlike disintegration, it is promoted by warm, moist climates, by moderate relief, and by a covering of vegetation. If the climate of Egypt were less dry, fewer pieces of old stone art would have been preserved for us; and if the climate of Labrador were less cold, a productive soil mantle might exist there.

The chief agencies that bring about decomposition of the rocks are oxygen and carbon dioxide from the air, water, the organic acids released by decaying vegetation, and certain bacteria.

Decomposition. — If a piece of iron is left exposed in a damp climate, it rusts; that is, it changes from iron, which is unstable in the presence of oxygen and water, and entering into combination with these substances, forms rust or limonite (iron, oxygen, and water), which is stable under the changed conditions. A simple chemical reaction takes place, and a new substance results. This is one of the changes that occur when rocks decay. The iron-bearing minerals of the rocks, for example, yield limonite, and the rock surface or resulting soil takes on the familiar tints of iron oxide, imparting to deeply weathered areas the red, yellow, or brown colors so pleasing to the eye.

But to produce fertile soil, other rock minerals must be decomposed. The chief rock-making minerals are the feldspars, and they are also the most important soil-yielding minerals. The chemical change of one of them, orthoclase, may be illustrated: —

Orthoclase+Water+Carbon dioxide yields Clay+Silica+Potassium carbonate
$[KAlSi_3O_8+H_2O+CO_2=\text{(Various hydrous aluminum silicates)}+SiO_2+K_2CO_3]$

This is one of the most important reactions in nature, since the existence of life so largely depends upon it. The necessary carbon dioxide is obtained from the atmosphere and in part from decaying vegetation, and it enters readily into solution with water to form carbonic acid. Clay is the basis of most good soils, and potassium carbonate is a necessary food of plant life. The potash chemically locked up in orthoclase is thus set free as a soluble food that can be assimilated by plants.

Similarly, other common rock-making minerals that contain aluminum, such as amphibole and pyroxene, yield clay. Some also yield potash. Quartz resists decomposition and remains in the soft decomposed materials as grains of quartz, or sand. When granite, composed chiefly of feldspar and quartz, is acted upon, the resulting soil is a mixture of clay and sand grains, called loam.

The combination of a substance with oxygen is *oxidation;* with water, *hydration;* and with carbon dioxide, *carbonatization;* all these processes enter into decomposition. As a result new substances called oxides, hydrates, or carbonates, are formed. Some remain behind in the soil, others are carried off in solution.

Solution. — In addition to the more complex chemical changes mentioned above, simple solution also aids chemical decomposition. Pure water (H_2O) is a poor solvent for rock minerals, but when it is combined with carbon dioxide (CO_2) to form carbonic acid (H_2CO_3) it is a powerful natural solvent. It attacks, among other substances, calcium carbonate ($CaCO_3$) and converts it into calcium bicarbonate [$H_2Ca(CO_3)_2$], which is quite soluble in water. Some rocks, such as limestone, are composed almost entirely of calcium carbonate, which forms the mineral calcite, and in others, such as sandstones, this substance acts as a cement to bind the individual sand grains together. When carbonic acid acts on such a rock, the binder is dissolved, the grains loosen, and the rock crumbles and breaks down into sandy soil.

Limestone may be almost wholly dissolved, leaving behind only the insoluble impurities, usually clay, to form a residual soil. Unbelievable thicknesses of rock have been removed by this process, at a rate that is possibly as great as 1 inch in 25 years. The impurities gradually accumulate on the surface, giving rise to the fertile soils of such limestone regions as southern Kentucky.

The impurities in the limestone are in some places iron or manganese compounds instead of clay, and residual accumulations of these substances have formed valuable deposits of iron or manganese ore that are being mined in different parts of the world.

Solution is also effective in removing some of the soluble products that result from chemical decay.

Plants and Animals. — Both plants and animals aid in the chemical attack upon rocks. Plants in their growth extract carbon dioxide from the atmosphere; they release oxygen and retain the carbon. On their death and decay some carbon dioxide is released and furnishes a source of carbonic acid for chemical attack upon the underlying rocks. As the plants die and others grow upon their ruins, a part of the carbon accumulates as a carbonaceous residue, known as *humus*, desirable as an ingredient of fertile soils.

The humus also supplies to the rain water that seeps through it small quantities of complex organic acids which are in themselves effective chemical agents of rock destruction. They supplement the work of carbonic acid in mineral decomposition or solution.

In addition, these acids are potent reducing agents; that is, they are able to take away oxygen from some oxidized compounds. This is most strikingly seen in the effect upon the ferric hydroxides that color the soils red, brown or yellow. The ferric hydroxides (Fe_2O_3 + water) are reduced to ferrous oxide (FeO), some of which unites with carbon dioxide to make ferrous carbonate, and this in turn is soluble in carbonic acid. Thus some of the iron is leached out and removed, and the remaining part no longer gives high colors to the rocks or soils. This is strikingly seen where red sandstones overlain by vegetable mold are grayish in hue just beneath; and rootlets that extend down into red rocks have a bleached zone surrounding them. Where ferric oxides, limonite or hematite, form the cement that binds the grains of sedimentary rocks together, its removal by the action of organic acids causes the rock to fall to pieces and so brings about the formation of soil. Where much organic matter is present in soils, vivid colors are absent. The two are incompatible. For example, in certain clays dark blue, dark gray, or greenish to black color denotes the presence of organic matter; and any iron present must be in the colorless ferrous form. But if these same clays are fired to make bricks, the organic matter is burned out, the ferrous iron is oxidized to the ferric state, and red brick results. If iron is not present, light-colored brick is formed.

Insects and other animals that live and move about in the soil aid chemical work by upturning the soil and exposing fresh surfaces to weathering. Their openings in the soil also enable weathering agents to reach more readily the underlying bedrock. Darwin states that in England earthworms bring to the surface 10 tons of mold to the acre every year, and Branner believes that in many tropical regions ants are even more effective in upturning the soil.

Bacteria also contribute their part. Certain forms have the unusual power of assimilating carbonate of ammonium and setting free nitric acid which attacks and decomposes rock minerals. They penetrate in great numbers every little nook and cranny in the soil or bare rock, and in time their effect is of no inconsiderable geological significance.

The general effect of chemical decomposition is to break up the complex rock minerals to form softer incoherent materials that either remain as a soil mantle, or are carried away by rain or wind.

Mechanical Effects of Chemical Weathering. — The impossibility of making a sharp separation between effects of disintegration and decomposition now becomes evident. Many chemical changes cause increase of volume. Hydration especially produces swelling of the parts of the rock affected and therefore sets up strains. The outer surface of an exposed rock dries rapidly; but moisture that penetrates into minute crevices remains until some decay is effected. Slow volume change at slight depth finally disrupts the rock, and pieces that are quite or nearly undecomposed are broken off. *Exfoliation*, the spalling

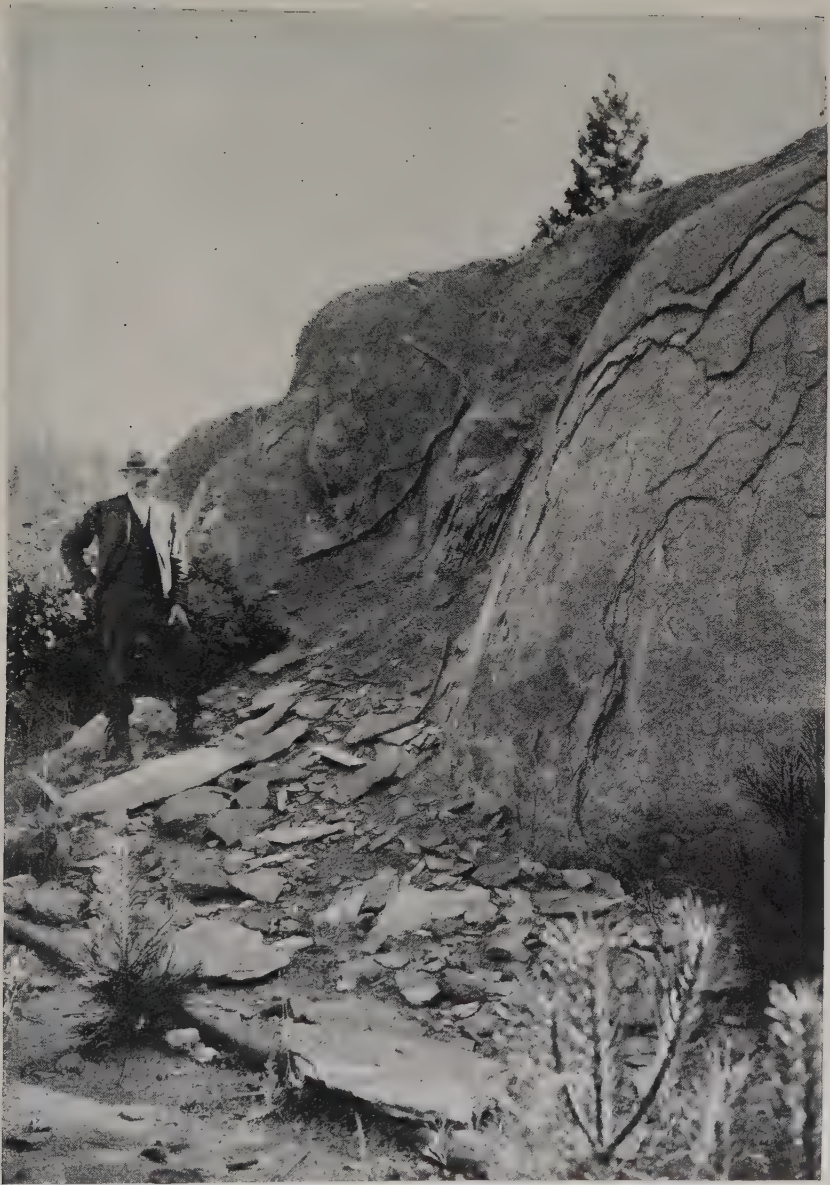


Fig. 12. — Exfoliation or scaling of rock. Commonly attributed to temperature changes alone, this scaling probably results in large measure from hydration of minerals, with consequent rupture due to increase of volume.

off of thin concentric shells (Fig. 12), probably is to be explained in this way, although some observers have attempted to explain it as due entirely to temperature changes. It is significant that exfoliation is common only in regions where moisture is fairly abundant, and not in deserts where rapid changes in temperature are most extreme. In arid lands a coarse-grained rock such as granite breaks up by disintegration into its constituent mineral grains. This effect may be due in part to temperature changes; but probably even in the driest regions slow hydration in crevices is an important factor in forcing the grains apart.

PRODUCTS OF WEATHERING

Soil and Rock Mantle. — Nearly everywhere the Earth's surface is mantled by a thin veneer of decayed or disintegrated rock to which the term *rock mantle* is applied. Its thickness varies considerably, but compared with the Earth as a whole it is but a film. In tropical regions it reaches a few hundred feet in depth; normally it is only a few feet, or a few tens of feet in thickness. The term *soil* is applied only to that part of the mantle which has been altered sufficiently to furnish the basis of plant growth.

Soils may be divided into two large groups: *residual soils,* which have been formed in place from the immediately underlying rock; and *transported soils,* which have been moved from their place of origin.

Residual soils normally pass gradually downward from a topsoil supporting vegetation into the subsoil of coarser material full of bits of rotted rock, and then imperceptibly into decayed rock that crumbles more or less easily; and this in turn merges into unaltered solid bedrock (Fig. 13). This gradual transition from topsoil above to solid rock below is one proof that the soil has been formed in place by the decomposition of the underlying rock. The transition between rock and soil is not only gradual but also highly irregular from place to place. Rich residual soils are widespread over most of the Piedmont district of the southeastern United States, and are the chief soils in humid tropical or subtropical regions.

Transported soils have been shifted by rain or running water (*alluvial*), by wind (*eolian*), by moving ice sheets (*glacial*), or by gravity down hill slopes (*colluvial*). Consequently they vary in composition more than residual soils and range from the finest silts to gravel. They have no essential kinship with the underlying rock since their source may have been far distant from their present site. Wide areas of the northeastern and north-central United States are mantled by glacial soils that

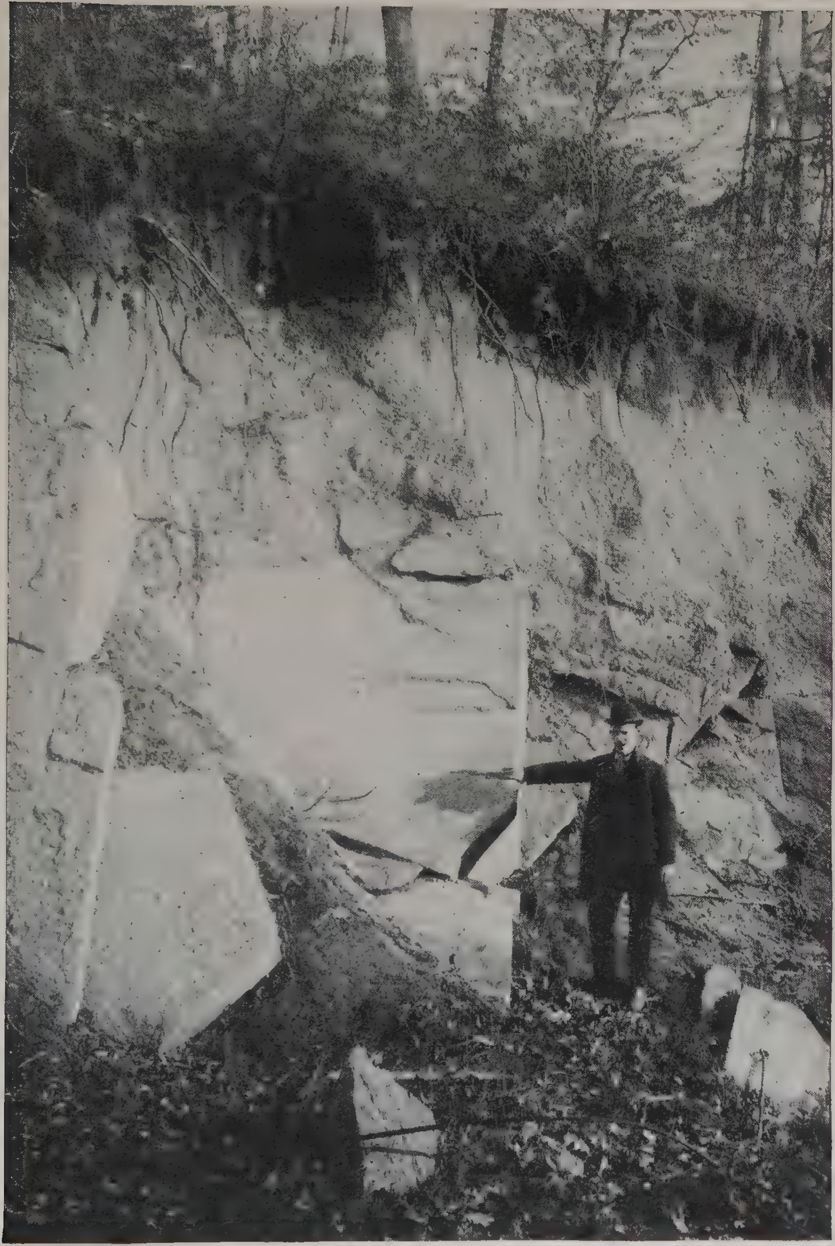


Fig. 13. — Residual soil formed by rock weathering and decay. The material graduates from firm rock below, through rotten rock and then subsoil, to true soil above. The transition is gradual. The soil above is colored dark by decayed organic matter. (Merrill, U. S. Nat. Mus.)

came from farther north, and the fertile silts that flank the lower Mississippi River came from far upstream.

Character of the Rock Mantle. — The character of rock mantle depends largely upon the kind of rock from which it was derived and upon the climatic conditions under which weathering took place. Thus rocks composed chiefly of feldspar under arid conditions disintegrate to feldspathic sand; under humid conditions they yield clay. A quartz sandstone yields only sand. According to the size of the particles in the rock mantle, the following gradations are recognized: Pieces of loose rock from the size of a small melon up are termed *boulders;* those larger than peas are called *gravel;* pieces smaller than peas, but not coherent when wet, are *sand;* and the finest material, which can be carried by the wind, is *dust;* the last is termed *silt* or *clay*, according to its character, and generally coheres when wet. Ordinary soils are composed of variable mixtures of sand and these finer materials.

Loam, a mixture of clay and sand, is easily worked and makes excellent soil; pure clay contains abundant plant food but is likely to be stiff and difficult to work; very sandy soils are usually unfertile.

Soils very rich in humus are called *muck.* When a clay soil contains a considerable quantity of calcium carbonate it is termed a *marl.* Thus *sands, loams, clays, mucks,* and *marls* are the chief kinds of soils and there are all gradations of these into one another. Owing to the presence of the dark organic matter or to the greater oxidation of the iron compounds, and often for other reasons, the topsoil is likely to be much more strongly colored than the underlying subsoil.

Talus. — In rugged country where rapid disintegration takes place by frost action, exfoliation, or otherwise, the rock so broken collects by gravity on the lower slopes, forming " slide rock " or *talus.* This is illustrated in Fig. 14. The talus either is a sheet-like form that flanks the slope or is cone-shaped, in which case it is called a *talus cone.* Some mountains are so buried in their own talus that only their summits project.

Residual Boulders. — The change of bedrock into soil is likely to proceed first along cracks and fissures. These usually intersect each other and inclose large and small blocks of bedrock. Consequently the agents of decomposition and disintegration begin their work on the outside of such blocks. The vulnerable corners succumb, then the sides are attacked; rounded *residual boulders* may result (Fig. 15). Where the attack of chemical and mechanical agencies has gone so far that rock masses or boulders take the shape of spheroids, it is called *spheroidal weathering.* Concentric layers, like the skins of an onion, continue to peel off until

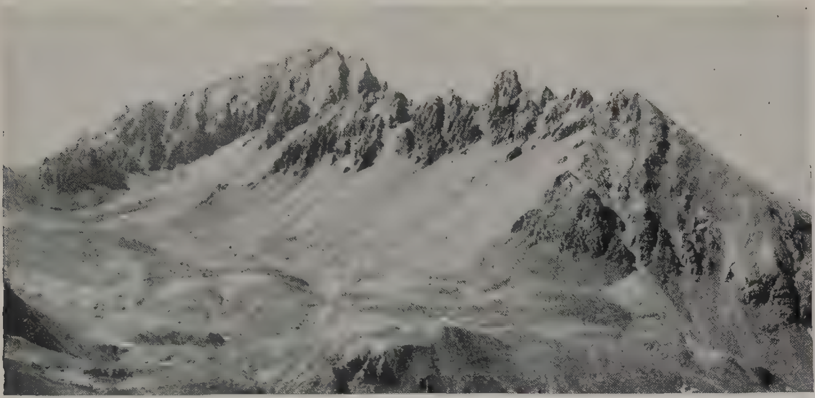


Fig. 14. — Rock disintegration and weathering in high altitudes with formation of long talus slopes of "slide rock." Mt. Sneffels, Colorado. (U. S. Geol. Surv.)

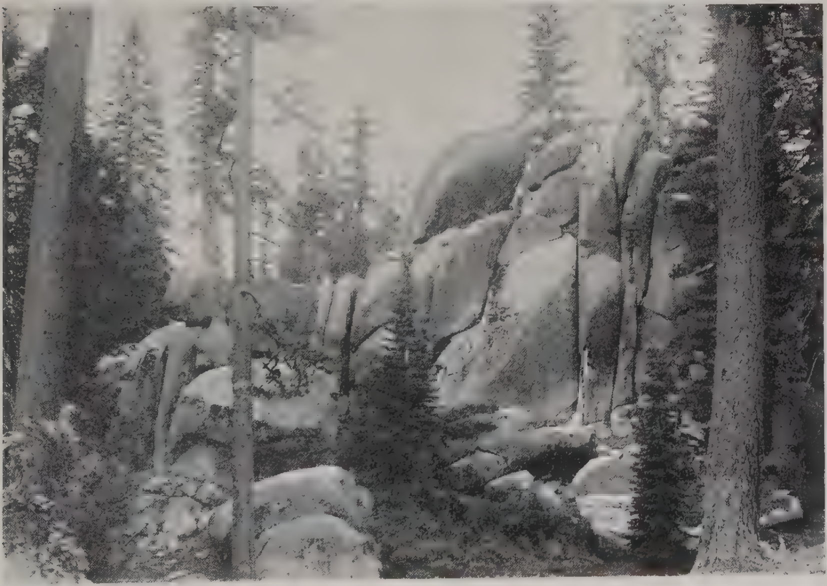


Fig. 15. — Jointed granite weathering into rounded residual boulders. (U. S. Geol. Surv.)

eventually the rock bodies disappear. Also, masses distributed through the bedrock are different in composition or texture from the rest, and thus harder or less soluble. Some of these also are left as residual boulders (Fig. 16). It must not be assumed, however, that all boulders are formed by this process, for many are transported blocks wholly unrelated to the underlying rock.

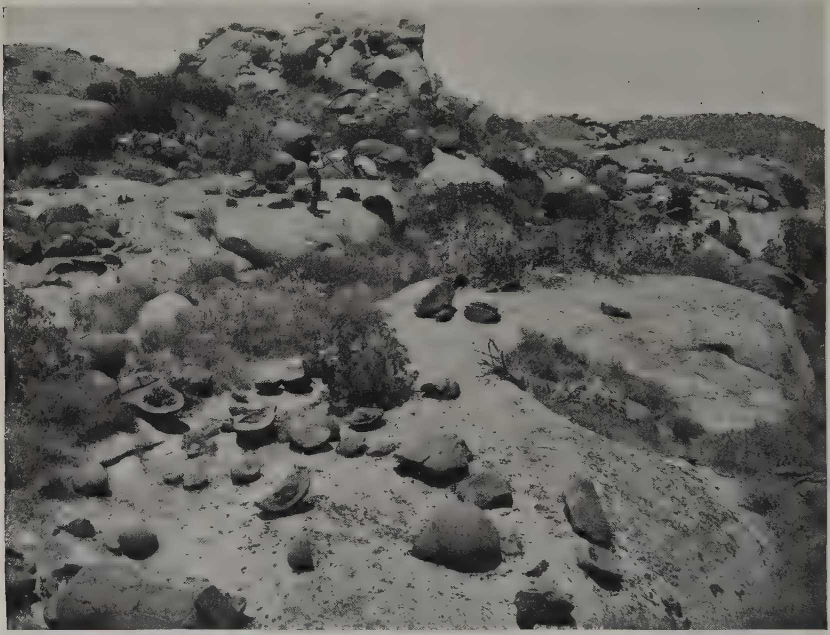


Fig. 16. — Residual boulders left by decomposition and wearing away of bedrock. The boulders are included masses of a harder, more resistant material and of rounded shapes (concretions). This shows that some residual boulders differ from the bedrock on which they lie. Coalinga, Calif. (U. S. Geol. Surv.)

WORK OF THE WIND

The weathering of rocks, discussed above, may be thought of as a static process, since it involves no transportation of the broken and decayed particles. The products of weathering would lie in place if they were not disturbed by other operations. But the atmosphere, in addition to disintegration and decomposition of rocks in place, plays also a more active rôle. The wind carries away products of weathering; it sweeps sand against rocks and abrades them; and it builds up dunes and dust deposits. Its work is most effective where vegetation is absent; consequently its results are to be seen in arid or semiarid lands and near sandy shores. The effect of the wind in forming waves that waste the coasts may best be deferred to the discussion of oceans.

The Wind by Itself. — The wind blows away products of weathering as they are formed, thus exposing fresh rock surfaces to further wasting. With its aid, therefore, disintegration and decomposition can proceed faster. This action proceeds best on hillsides or valley slopes where lodgment of weathered particles is insecure and where gravity aids; but it also operates even in flattish areas, particularly in arid regions, where mechanical agencies disintegrate the surface, and wind blows the material away, making shallow depressions. This blowing away of loose material is termed *deflation.* In semiarid regions such as Nebraska, in times of drought, the wind dries the soil and blows it away from cultivated fields, with disastrous results to young crops and the supply of soil.

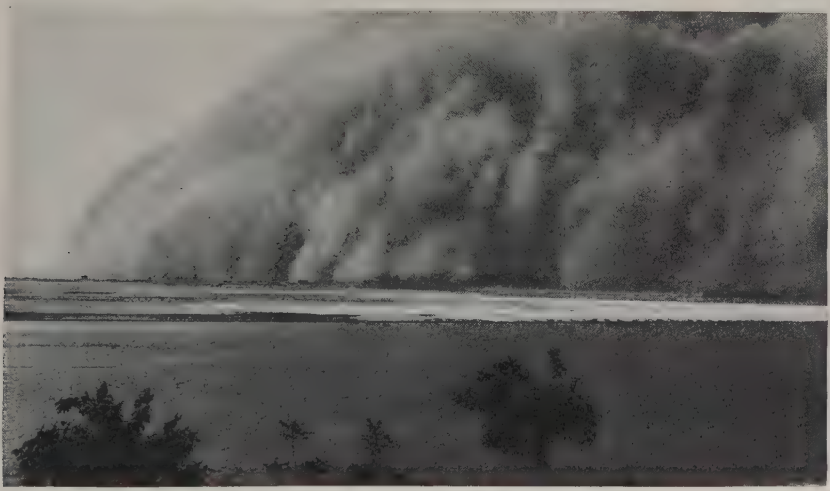


Fig. 17. — Sand storm sweeping over Kartoum North; in front is the Blue Nile. Shows enormous transporting power of the wind. Soudan, June 6, 1906. (Wm. Beam.)

The wind transports vast quantities of material. A gentle breeze lifts and carries dust, a strong wind drives sand (Fig. 17), and a tempest can move gravel the size of peas. A single storm that travelled from the arid Southwest a thousand miles to the Great Lakes region brought with it a million tons of dust, which was deposited on the snow over a wide area. Such material, dropped when the wind slackens, in some places eventually forms deposits of great magnitude. They are known as *eolian* (Aeolus, god of the winds) deposits, to distinguish them from water-laid deposits.

Abrasion by Wind-blown Sand. — The smoothing effect of sandpaper applied to a rough surface is a familiar example of abrasion. The particles of sand fastened to the paper make sharp and effective cutting tools. Similarly, begrimed stone or brick buildings today are being renovated by a sandblast — sand blown by compressed air. Nature supplies a similar agent of abrasion in the sand moved by wind. Though less efficient as a cutting tool than its artificial counterpart, it is, over long periods of time, an important agent of rock destruction.

The rate of cutting by storm-driven sand is locally quite rapid. For example, window panes of houses along seashores have lost their transparency in a day or two, and have been completely penetrated in a month or so. Wooden telegraph poles planted in the desert are in some localities quickly cut down by the sand drifting past their bases, and on stretches of railroad that cross deserts, rails have to be replaced frequently because of abrasion by drifting sand.

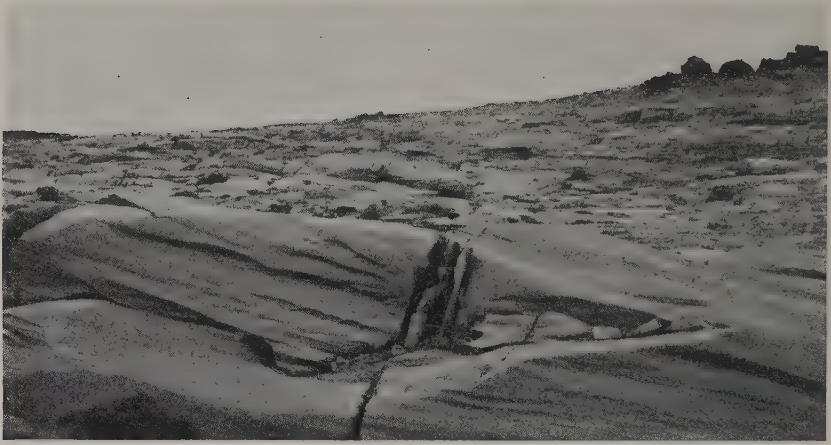


Fig. 18. — Limestone polished and grooved by wind-blown sand. Namib, Southwest Africa. (A. W. Rogers.)

The most effective abrasion by wind-driven sand takes place in deserts, where rainfall is scanty and protective vegetation is sparse. Exposed rock masses are carved and sculptured into rounded (Fig. 18) and " hoodoo " forms. In detail, the surfaces are commonly pockmarked or intricately *fretted* by the removal of softer parts. Further evidence of eolian abrasion is to be seen in the smoothed, polished pebbles that lie in such areas. Many of them are faceted into angular forms, commonly with triangular cross sections. Such pebbles are called *dreikanter*.

The wearing away of the land through the combined effects of abrasion and ablation is termed *wind erosion.* Erosion is a general term applied to the wearing action of several agencies. The wearing effects of running water, glacial ice, and waves will be considered in later chapters.

Sand Dunes. — Sand dunes are hillocks made by wind-borne sand in a manner similar to that in which snow forms drifts. Their growth is started by some obstacle, such as a rock, a bush, or a surface irregularity, which gives local protection from the wind or causes wind eddies, so that the moving sand is deposited. Even the erection of buildings starts dune formation in some places. Commonly the surface of a dune is covered with fine parallel ridges of sand an inch or so in height, transverse to the direction of the prevailing wind, and called *ripple marks* because they resemble the markings on sand made by water ripples (Fig. 19). Dunes are formed wherever free-moving sand exists and where there are prevailing winds. Frequent changes in wind, as well as the presence of thick vegetation or of exceptionally rough topography, are unfavorable to their development. They are found in all parts of the world along low coasts where sand made by waves is washed ashore and, caught up by the prevailing winds, is drifted inland and accumulated. Dunes so formed occur at various places along the coasts of the United States; in England; on the shores of the Baltic Sea; in Holland, Belgium, and France. Similarly they are produced on the shores of large lakes or inland seas; the southern end of Lake Michigan is fringed with high sand dunes. They are also of common

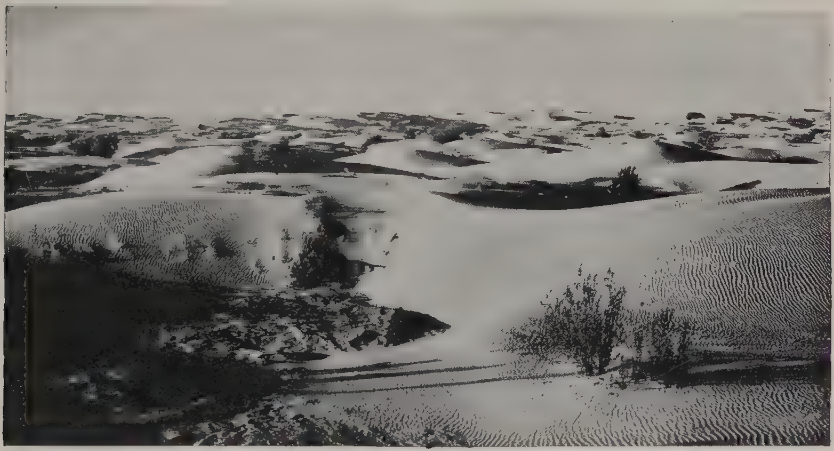


Fig. 19. — Sand dunes near Mammoth Station, Calif., showing ripple marks. (U. S. Geol. Surv.)

occurrence along broad river bottom lands in semiarid regions, as along the Arkansas River in western Kansas. Sand dunes are especially abundant in arid regions, where disintegration is rapid and a protective mantle of vegetation is lacking. It is a mistake, however, to suppose that all desert lands are mantled deeply with sand. Even in the Sahara there are great areas of barren rock from which the wind sweeps all loose material.

Shape of Dunes. — The shapes of dunes vary according to the intensity and direction of the wind. Some are linear, and others very irregular; many are curved or crescent-shaped in plan and are called *barchanes.* The windward side of a dune has a gentle slope which is determined by the strength of the wind. The leeward side, capped by a rather sharp crest, has a steeper slope, which is the angle of repose of free falling sand, and varies from about 20° to 30°. The shape thus tells the direction of the prevailing wind.

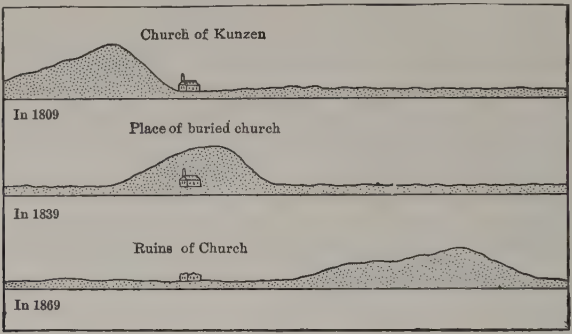


Fig. 20. — Movement of a sand dune during 60 years on the east shore of the Baltic Sea at the village of Kunzen. (After Berendt.)

Migration of Dunes. — A dune once formed is never stationary unless it meets an object larger than itself, or is held by a mat of vegetation. It moves ceaselessly onward with the prevailing wind, by the slow transference of sand grains from the windward to the leeward side. During this progress its height is maintained or increased, though its outline may suffer change. Thus, along some coasts where the prevailing winds are from the sea dunes form a belt extending several miles inland. Their rate of movement may average about 20 feet a year, as in Denmark, or reach more than 100 feet a year, as on the Biscayan coast of France. In their march they cover and destroy arable lands, forests, and even towns, leaving desolate sandy wastes behind (Figs. 20, 21). In the deserts of central Asia, Sven Hedin, the explorer, found ruined cities of an ancient civilization emerging from sand dunes which for a long time had overwhelmed them and the fertile lands that once supported them.

The devastating migration of dunes has been mastered along the coasts of European countries by the skillful planting of trees or other vegetation to prevent the removal of the sand particles from the windward slope. The dunes then become stationary. In France the planting of forests on the dunes resulted not only in preventing dune migration but in the raising of profitable forests on otherwise waste land. In arid regions, however, where vegetation will not grow, the migration of dunes is almost impossible to control, and nothing should be built in the path of their travel.

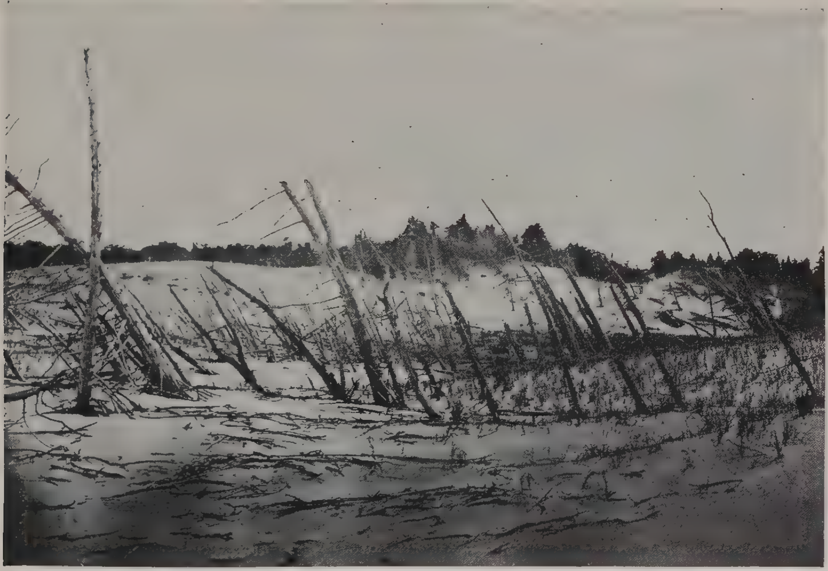


Fig. 21. — Forest overwhelmed, killed, and then left exposed by marching sand dunes. Manitou Island, Lake Michigan. (U. S. Geol. Surv.)

Loess. — Extensive deposits of fine clay-like materials have been built up chiefly by the wind to form *loess*. The name comes from Alsace, where it designates a peculiar fine-grained, yellowish-brown, lightly compacted earth. It is remarkably fertile. The deposits show no horizontal banding, but form upright bluffs because of a tendency to cleave vertically (Fig. 22). These deposits are extensive along the Rhine, occurring well up on the mountain slopes, and form some of the most fertile soils of central Europe. They cover tens of thousands of square miles in the central Mississippi Valley region, especially in

Iowa, Nebraska, and Kansas, and are found also in Oregon and Washington. Similar extensive loess deposits occur in the rich pampas of the Argentine. They are thick deposits that have formed chiefly from accumulations of dust, which in America and Europe was probably supplied by the flood plains of glacial streams during the recent ice age. There was no vegetation to hold the soil at that time, and the wind blew away the finer particles.

The greatest development of loess is in Asia, particularly north central China, where it covers 230,000 square miles, and the yellow earth washed down from it has given the Yellow River and the Yellow Sea their names.

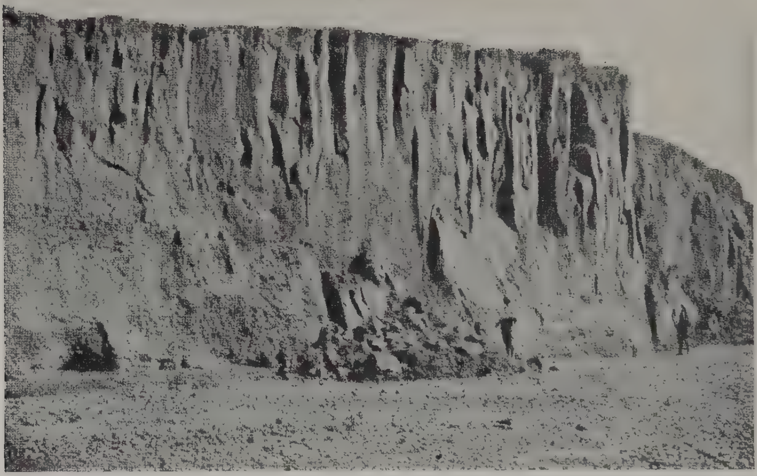


Fig. 22. — Loess lying above gravel, Wu-tung, China. Note the vertical jointing, which keeps the bluff vertical, although the soft loess is easily eroded. (Barbour.)

It is believed by Von Richthofen to have been carried by the wind from the great deserts of the interior, especially the Gobi desert, and to have spread over the adjacent area. Its thickness reaches 300 feet or more. It is remarkably fertile, and even though it has been cultivated for thousands of years, it is still productive. Streams have cut canyons into it, and at the base of the bluffs the humbler Chinese have fashioned in it cave dwellings which they have used for centuries. Where roads cross it, the material loosened by the cart wheels has been blown away, and in the course of centuries this process has caused the roadways to sink lower and lower, so that some of them are now small canyons.

CHAPTER IV

RAIN AND RUNNING WATER

The Rainfall. — There are very few parts of the Earth's surface, if indeed there are any, on which rain does not sometimes fall. Death Valley, in California, one of the driest places in North America, has about 2 inches of rainfall annually. Even the most arid spots in the Libyan Desert have had rain at least once within a measured span of 12 years. But the amount of rainfall a country receives is dependent on a variety of factors, such as the direction of the prevailing winds, the nature of the places over which they have previously passed, and the height above the sea of the country which receives them. Thus it happens that the amount of rainfall received by the land is very unequally distributed over the world; in many places it is as much as 100 inches per year and in parts of India 500, but in the great deserts it is less than 10. Roughly speaking one may term as *arid* those regions where the rainfall is less than 10 inches, as *semiarid* those where it is 10 to 20 inches, and as *humid* those where it is more than 20 inches (Fig. 7).

After the rainwater reaches the surface of the land part of it is trapped by surface depressions and remains in lakes and ponds. A part evaporates, a part sinks below the surface and becomes ground water, and a part flows down the slope of the surface, forming the *immediate run-off*.

The Run-off. — The proportion of the rainfall that contributes to the run-off in a given region depends on several factors such as (1) surface slope, (2) porosity and solubility of surface material, (3) character and amount of vegetation, (4) temperature and humidity of the atmosphere, and (5) distribution of the rainfall throughout the year. It therefore varies greatly from place to place. Certain areas of very soluble and porous limestones have essentially no run-off, all of the rainfall sinking below the surface, whereas on steep slopes in the Appalachian Mountains, nearly all of the water from rains in the early spring flows off over the surface, because the ground is already saturated with melted snow. On an average, about one-fifth of all the rain that falls forms run-off. Since much of this fifth flows from areas thousands of feet high to the sea, the power of the resulting streams to do work is very great. It is estimated that all the streams of the United States develop, on the average, 419,000,000 horse power. As the water under the influence of gravity flows off over the surface during and after rains, it seeks the shortest and steepest route downward, following the lowest depressions wherever they present themselves, and is continuously augmented by seeps and springs representing the reappearance of the ground water which sinks below the surface at higher levels.

Stream Valleys; Stream Erosion. — Every observer, as he stands on the brink of the Grand Canyon of the Colorado and looks down at the winding ribbon of the river a mile below, is impressed with the vast size of the rock-walled valley and with the puny appearance of the stream at its bottom. The earliest observers thought that the Colorado had found this low course already prepared for it by a great gash or rift in the Earth's crust. But the visitor of today, if he knows some of the principles of geology and understands rivers and their behavior, can believe that the river once flowed through a shallow trough at the level of the canyon's brink, and that very gradually the sediment-laden water cut down little by little into the underlying rocks, sawing ever more deeply until in the course of time it dug a trench a mile deep. This same visitor surveys from an eminence the intricate pattern of the canyons and sub-canyons which empty into the main valley, arranged symmetrically like the veins in a leaf, most of them entering flush with the main stream. He realizes that they are not fortuitous, but are all parts of a unified and interrelated system developed in obedience to a common law, and even having a common destiny. And when he sees this same pattern repeated again and again in stream systems large and small throughout the world, he concludes that all rivers act according to certain definite rules imposed on them by their surroundings. The fundamental cause of stream flow is the pull of gravity upon mobile liquid at the Earth's surface. But as water flows it picks up loose rock particles, carries them downstream, and by their aid wears away the solid rock. This process is called *stream erosion.*

Consequent Streams. — Consider the ideal history of a land area newly emerged from the sea. During the first rainfall following emergence, the impact of raindrops upon the surface loosens the exposed fine particles of rock mantle and the run-off carries them down the nearest slopes in little temporary rills or rivulets. Where rivulets join at the bases of converging slopes the wash of their combined volumes digs out *gullies.* Adjacent gullies join at lower levels and the constantly increasing volume of water excavates *ravines.* These in turn join to form even larger *valleys.* In this way, with many more junctions and convergences, the run-off reaches the sea in the form of streams loaded with sediment picked up at higher levels. The slopes down which the run-off flows in this early stage are called *initial slopes,* and the streams following or "consequent upon" these slopes are known as *consequent* streams. The rate of flow of the consequent streams depends entirely upon the steepness of the initial slopes.

The rivers are the main channels of drainage, and they are the chief factors in carrying away the waste of the land. They are the great trunk lines of transportation for the rock débris delivered to them by their tributaries. In addition they are themselves powerfully capable of wearing away the land through which they flow; in them the work of running water as a geological agency is most conspicuously displayed. If we should think of a typical river we should imagine it rising in lofty mountains through the union of many impetuous streams or dashing torrents; gathering headway it rolls rapidly through the belt of lower hilly country and emerges upon wide plains through which it wanders in many curves in a quiet and steady flow to the sea. The slope of its bed, its *gradient*, which may be as much as 20° to 30° at the head, becomes less and less until it is nearly horizontal at the river's mouth (Fig. 23).

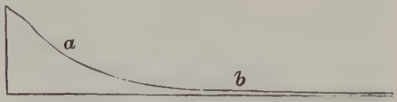


Fig. 23. — Profile of a normal river showing its gradient.

Although we think of this as the ideal course of a river (and it is typical of many of the great rivers of the world such as the Amazon and the Ganges, and of a great number of smaller ones such as the Po in Italy), we constantly find variations from this type. Thus the Mississippi does not rise in a mountainous country, but in a moderately elevated region of low relief, and except for the falls at Minneapolis the gradient is low throughout the entire course. On the other hand, some rivers rise in mountains near the sea, and hence their lower plains districts, corresponding to *b* of Fig. 23, may be short or wanting. In North America, the rivers of the northern Atlantic coast are mostly between these extremes and their courses lie between *a* and *b*; those of the southeastern states are more nearly typical, since they rise in the Appalachian Mountains and flow out upon the Atlantic coastal plain.

ANALYSIS OF STREAM EROSION

Erosion by streams consists of three distinct processes, (1) *corrasion*, (2) *solution*, and (3) *transportation*. The static process of weathering (Chapter III) precedes and accompanies erosion, preparing rock waste for easy seizure by running water, and deposition always follows it. But these two processes are entirely distinct from erosion proper.

1. Corrasion

Stream erosion is limited in its operation, being confined to the bottoms and sides of the channels through which the water passes. A river may be compared to a sinuous, flexible, and endless file, ever moving forward in one direction, and by means of the moving sand or gravel rasping away the rock beneath and beside it, thus cutting an ever-deepening trench. This particular phase of a river's work is called *corrasion.* The effectiveness with which a river corrades depends on several closely related factors; (*a*) on the abundance and character of the tools with which the river has to work, (*b*) on the velocity of its current, and (*c*) on the nature of the bedrock with which it has to deal.

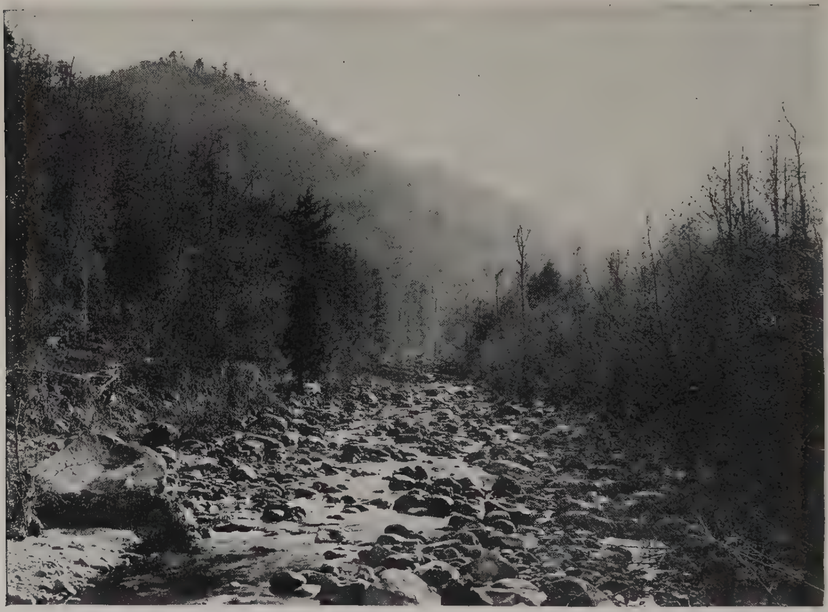


Fig. 24. — River bed full of more or less rounded boulders, showing the tools with which the stream works. Big Creek, Haywood County, N. C. (U. S. Geol. Surv.)

(*a*) **The River's Tools.** — Clear water moving over rock surfaces has little erosive effect. It has a certain solvent power and may thus slowly dissolve and disintegrate rocks, but in order to corrade, a river must have tools. These are supplied by the sand and silt which it carries, and by the gravel and pebbles it can move if swift enough, either in its normal flow, or in times of flood (Fig. 24). This material is supplied to the river chiefly by rain wash and by its tributaries, but the stream also obtains it directly by wearing and undermining the sides of its channel. If its banks are steep, or cliff-like, the débris which accumulates at the foot of such slopes is seized by the river and used as its tools. It is by the striking, bumping, and grinding action of this material, carried along by the current, that the river is able to cut away the rocks over which it runs and thus to deepen its channel.

In this process the material carried by the river is itself necessarily worn, has its sharp angles removed and becomes rounded or spheroidal, — a form characteristic of the river's tools. Thus, if we find river gravel which consists of hard, well-rounded pebbles, we infer the material has been transported a long distance; on the other hand if the gravel is composed of angular bits of rock, and its situation shows that it has been transported, we infer that the distance must have been short.

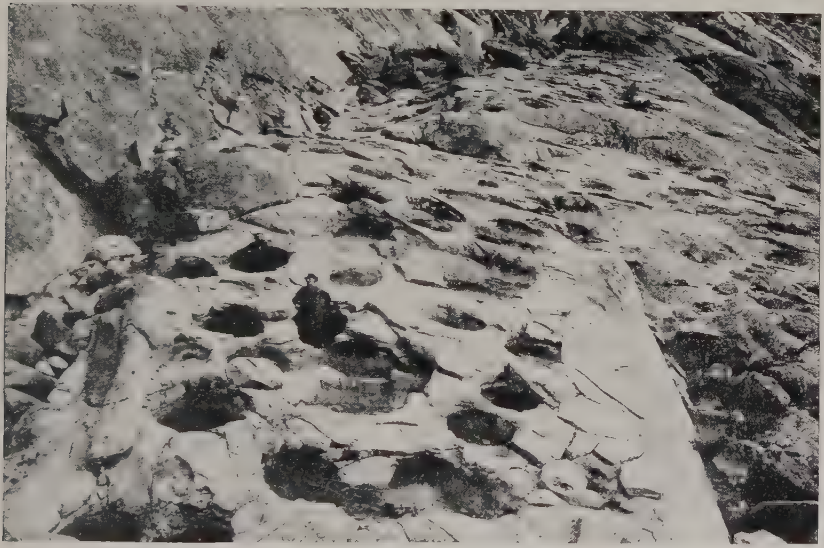


Fig. 25. — Potholes in granite. Tuolumne River, Calif. (U. S. Geol. Surv.)

The particular way in which a river uses its tools, and details of the effects produced, vary greatly. In a stream flowing over horizontal layers of rock, corrasion along joints loosens large blocks, which are then torn or "plucked" away by the current. In some situations this plucking action is much more important than wear by simple rasping. As each layer is stripped back by the process, the bed of the stream exhibits a succession of low steps, descending abruptly from one layer to the next lower.

Circular excavations worn in bedrock by whirling eddies are common in the beds of streams below falls and rapids. They are called *potholes* (Fig. 25). If the conformation of the stream bed is such that an eddy persists in one place, the water whirls sand and gravel with it, and this bores downward; although the material wears out in grinding, it is continually replaced by fresh débris, and so the process continues. Potholes have diameters ranging from a few inches up to 50 feet; their depth may be even greater. They are of interest in that they indicate clearly the action of whirling water and, occurring not uncommonly in rock now far from any stream, they prove that at one time this rock was the bed of a rapid current.

Up to a certain point an increase in the amount of grinding material supplied to a river with a given velocity of current aids in its corrasive power. Beyond this point an increase is not effective for the reason that the strength of the current is so consumed in the operation of transporting that the check, given by a tendency to corrade, would cause the river to deposit instead of carrying material farther. Since corrasive power depends on the strength of the blows struck by the moving particles, it is clear that this in turn depends upon the momentum, which is determined by their mass and the velocity with which they move. Hence for a constant velocity, the greater the mass of the particles — that is, the larger and heavier they may be — the more effective agents of erosion they become. Thus in a stream carrying intermingled grains of sand and dust-like particles the sand is the more effective agent.

(*b*) **Velocity and Corrasive Power.** — It is obvious from the preceding paragraph that other things being equal, the swifter a current is the more rapidly it will corrade. For, in a given time, not only will the number of corrading particles passing over a rock surface be increased with a swifter current, but the fact that each particle is moving more rapidly will add to its effectiveness. Further, since a swift stream can carry larger particles than a slower one, and since these larger particles, owing to their momentum, strike the channel with greater force, it appears that a moderate increase in the velocity of a stream will result in a great increase in that stream's cutting power. Calculation of these complex factors shows that doubling a stream's velocity increases its corrasive power at least four times, and in extreme cases many times more.

(*c*) **Character of the Bedrock.** — Such rivers as the Platte and the Missouri on the Great Plains, the James, the Roanoke, and the Savannah on the Atlantic Coastal Plain, the Yangtse on the coastal plain of China, the Thames in the London Basin, the Seine in the Paris Basin, and a host of others besides — all are turbid with sediment. They are alike in that they flow through regions of shales, clays, and soft sandstones. On the other hand most of the streams of New England and the Adirondacks, and of the Highlands of Scotland, together with such streams as the upper Danube, are relatively clear. And these are alike in that they drain regions of hard igneous and metamorphic rocks. Thus we find countless illustrations of the fact that the composition of the rock in a stream valley has a great influence on the rate of erosion. The structure of the rock masses also exercises a profound effect. If they are jointed and thinly bedded, many planes of weakness are present along which the stream can chip out and dislodge fragments. Vertical beds of shale present ideal conditions for this kind of erosion. If the rocks are massive, like granite, streams traversing them erode with difficulty.

2. Solution

Even without visible tools running water wears away the land by solution. Its solvent action is greatly aided by substances already in solution, inherited in part from the ground water that issues upstream in springs and seeps and in part directly from the rainwater which commonly contains dissolved gases acquired from the atmosphere and from decaying vegetation. The rock most susceptible to the solvent action of stream water is limestone (calcium carbonate). But even where a stream flows over relatively insoluble rocks, certain soluble minerals are decomposed, thus loosening the adjacent insoluble mineral grains and preparing them for mechanical seizure by the current.

3. Transportation; the River's Load

The material carried by a stream forms its *load*. Although the greater part of this is carried (*a*) mechanically, either in suspension or by rolling on the bottom, a very considerable portion is transported (*b*) chemically in solution. The ultimate goal of the river is the sea into which, at the end of its journey, its remaining load is deposited. The various aspects of this work demand consideration.

(*a*) **Material Carried Mechanically.** — The size of the particles that a river is able to carry depends on (1) the character of the river's current, (2) its velocity, and (3) the relative weight or specific gravity of the particles.

1. *Character of the Current.* — If the mass of water forming the current moved forward in a perfectly uniform manner, each particle of water from side to side and from top to bottom moving forward with the same velocity as every other particle, only the very finest material, such as microscopic granules of clay, would remain any length of time in suspension. A sand grain dropped into the stream would sink to the

bottom and there remain at rest, unless the stream were strong enough to roll it along. But the current of streams is not of this character. The more central portions are moving more swiftly, sliding over those toward the bottom and sides, while there is a constant interweaving of swifter sub-currents up and down and toward the sides and even backward, forming eddies or whirling movements. The whole effect is like the stirring of water in a glass. Sand at the bottom is quickly lifted and kept in suspension by these movements while it is carried along by the main current. When particles in suspension in water attain a certain degree of fineness their settling, even when the water is still, becomes very slow. Thus water such as that of the Mississippi may remain turbid for many years.

In addition to the material carried in suspension a considerable part of the river's load is pushed or rolled along the bottom. What proportion of the whole this may be cannot be accurately determined; it has been thought that in the case of some rivers it is greater than the amount carried in suspension. From studies made on the Mississippi, it is roughly inferred that of the material which it carries into the Gulf of Mexico about 10 per cent or more consists of coarser débris moved along the bottom. It is obvious that, other things being equal, the steeper the gradient a river has the larger will be the amount of the material so moved.

By observation of experiments in troughs it has been found that the amount of material moved by actual sliding or rolling of the particles along the bottom of a stream is much smaller than the amount which progresses by a series of short leaps. Near the bottom of every stream there is a zone filled with grains moving in this fashion (*saltation* = jumping), and above this is the material in suspension. The particles urged forward by sliding, rolling, and saltation are said to be moved by stream *traction*. The amount carried by traction, compared with suspension, varies with the swiftness of the current and size of the grains.

2. *Velocity and Transportation.* — The velocity of a current depends not only on the gradient, but also on the volume of water involved. Thus of two streams having similar gradients and form of channel, the one having the larger volume of water will have the swifter current. It is also well known that the swifter a current, the larger and heavier the masses it can transport. It has been found that a current running a fifth of a mile in an hour will carry fine clay; one running half a mile in an hour will transport sand; one of a mile an hour will roll along medium-sized gravel, and one of 2 miles an hour will sweep along pebbles the size of an egg. Reduced to mathematical form it may be stated

that *the maximum size of particles transported by a current varies as the sixth power of the velocity.* In other words, if the velocity of a stream be doubled it can move particles sixty-four times as large as it moved before. With low velocities of less than a mile an hour and with fine particles this increase does not seem very striking. But with increasing speeds of miles per hour the effect becomes very marked. This explains why rapid streams of 5 miles per hour are able to move small boulders, and sudden floods in narrow valleys, caused by torrential downpours of rain or the bursting of dams, are able to carry with them huge masses of rock, sweep away bridges and other structures, and cause great damage (Fig. 26). When the St. Francis dam near Los Angeles gave way in 1928 and flooded the valley below, huge blocks of concrete, one of them weighing 10,000 tons, were displaced by the escaping waters.

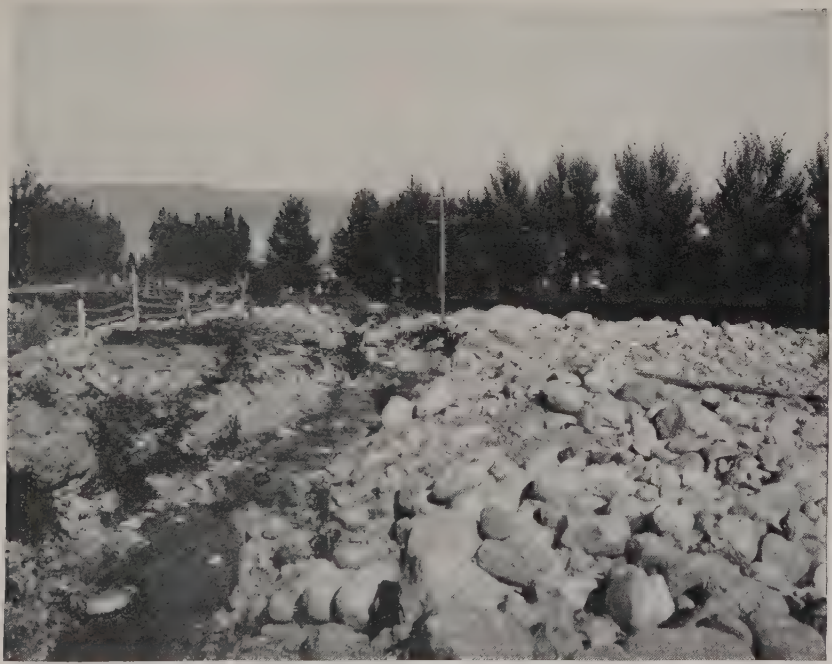


Fig. 26. — In times of flood a stream is able to carry masses and perform work vastly greater than under ordinary conditions, as here shown by the results of flooding. Manti Creek, Utah. (U. S. Geol. Surv.)

It should be kept in mind that this law of the sixth power applies to the maximum size of particles and not to the total load that a stream can carry. The largest masses are of course moved by rolling along the bottom. The total amount carried by traction and suspension increases greatly with velocity, but so many complex factors are involved that no precise statement of the relation can be made.

3. *Effect of Specific Gravity.* — The size of the particle that a stream of a given velocity is able to carry depends also on the specific gravity of the materials composing the particle. A familiar example is that a lead sinker is able to remain at rest on the bottom of a stream which carries away pebbles of an equal size. A practical application of importance is found in the fact that in placer gravels, from which gold is extracted, fine particles of the precious metal are mixed with vastly larger ones of sand and gravel; the water, on account of the high specific gravity of the gold, being able to transport it only with difficulty. When a prospector finds angular grains of gold, he searches the adjacent slopes for gold-bearing veins. The specific gravity of the great mass of material obtained by erosion and carried by rivers lies between 2.5 and 3.0; that is, it is that much heavier than an equal volume of water. The principle that a body immersed in water loses weight equal to that of the water displaced, greatly aids the transporting power of the stream.

(*b*) **Transportation of Material in Solution.** — All river waters carry in solution salts of various kinds that have been leached from the rocks and soils of the country they drain. Although in a measured volume of what we call fresh water the amount may seem relatively very small, in the aggregate the weight of material thus dissolved from the land and carried into the sea is enormous. It has been estimated that nearly 2,735,000,000 metric tons[1] of solid substances are thus annually transported into the oceans. The Mississippi carries about 136,000,000 tons, the Connecticut, a small river, 1,000,000, the Danube over 22,000,000, the Nile nearly 17,000,000. In the Mississippi the amount carried in solution is more than a third as large as that carried in mechanical suspension, the quantities being

340,500,000 tons in suspension;
136,400,000 tons in solution.

The most important of the substances thus dissolved and transported are calcium and magnesium carbonates; calcium, sodium, and potassium sulphates; sodium chloride; and silica.

Manner of Transportation. — In considering the manner in which material is carried one must recall that it is only in swift streams and the upper rapid tributaries of great rivers that boulders and coarse gravel are moved, generally in times of flood. As one goes down a great river the size of the material steadily grows less with diminishing gradient. This is seen not only in the matter in suspension, but on the bars and

[1] Metric ton = 1000 kilograms = 2204 pounds.

beaches where it is temporarily deposited. Finally, in those rivers which wander through wide plains before they reach the sea, only the finest sands, silts, and clays are discharged into the ocean, and no coarse material is seen, except chance pebbles and boulders that have been floated downstream in masses of river ice or among the roots of drifting trees.

Nor is the journey a steady or uninterrupted one. The gradient changes from place to place and with it the velocity and transporting power. Material carried down one reach is deposited at the foot of it, while at the head of the next, rapid erosion is cutting the channel headward and material is thus again set in motion. Matter dropped during a season when the current is slack is seized and again hurried forward with the renewed strength that comes in times of flood. Thus, with many waits and pauses, and growing finer by attrition, the mass of material upon which the river works is urged ever forward and onward down stream.

Rate of Erosion; Denudation

It has been estimated that the amount of material in suspension, in solution, and rolled on the bottom, discharged each year into the Gulf of Mexico, if gathered together would form a right-angled prism with a base 1 mile square and a height of 250 feet. If we reckon the whole basin of the Mississippi and its tributaries as covering 1,265,000 square miles, and consider only the material in suspension and solution, it can be calculated from the given data that the entire basin is being lowered at the average rate of 1 foot in 6000 years. An estimate for the whole United States, based on measurements made on its rivers, is about 1 foot in 9000 years. The actual rate is probably greater, because the amount moved by traction is not included; and the two rivers, the Mississippi and the Colorado, which together transport about 80 per cent of the total material taken from the United States each year and delivered in suspension into the sea, certainly move a large amount by traction also.

The gradual lowering of a land surface through a long period of time is referred to as *denudation.* From the foregoing statements it is obvious that we cannot estimate the rate of denudation with any accuracy, but the results are of interest and importance because they indicate the order of magnitude of the figures concerned. We may say, with some confidence, that the area of the United States is being lowered at a rate of 1 foot in from 5000 to 10,000 years, and probably between 7000 and 9000; and when we consider that during past ages rock many thousands of feet thick has been removed by denudation, we get some notion of the immensely long periods of time involved in the process.

Estimates based on other rivers have given different figures. Thus it has been calculated that the Ganges erodes its basin at the rate of 1 foot in about 1750 years. But its basin culminates against the loftiest mountains in the world and the river has a proportionately rapid descent and great erosive power. The basin is also subject during part of the year to a very heavy rainfall and great floods. Consequently the rate is far greater than the average. On the other hand desert regions, like those in central Asia or the Sahara in Africa, with very little rainfall, are eroded with great slowness, the chief agent of transport being the wind.

STREAM VALLEYS

Development of Valleys. — Any land surface on which rain falls sheds the run-off down the lowest routes to the sea. The run-off loosens the surface soil along these low channels, carries away the débris, and thus excavates gullies. And stream valleys are merely gullies grown big. The evolution of gullies into valleys takes place in this way:

Gullies form where the run-off is concentrated. Concentration of run-off enlarges a gully by erosion on its side slopes and at its head. The larger the gully becomes, the more thoroughly it concentrates the run-off, and an endless chain of cause and effect is thereby set up. Erosion at its head causes the gully to lengthen headward, and the slope wash down its sides widens it after each successive rain. At the same time the ever more concentrated run-off through the bottom deepens the gully and lengthens it mouthward as well. Thus constantly undergoing enlargement, the gully becomes first a ravine and then a valley. The normal cross section of a valley which is undergoing rapid erosion is that of a V (Fig. 27), because the river, occupying a relatively small space, is cutting downward, while at the same time rain wash and gullying tend to broaden the trench the river makes, by washing down the material composing the valley walls. The cross section of a valley depends then on the relative balance between two agencies, downcutting by the river and broadening by weathering and rain wash. Thus in a region where the gradient is steep, downcutting by the river proceeds much more rapidly than weathering and rain wash, and the valley will be deeply incised and have profiles approaching *ara* (Fig. 27). As time goes on and the river gradient is lessened the cutting by the river

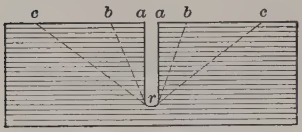


Fig. 27. — Section of a river valley; *ara*, material removed by corrasion; *arb*, *brc*, material removed by weathering and rain wash; river *r*, trenching downward.

becomes slower and slower; weathering then becomes relatively more and more pronounced and the valley widens out as shown in the diagram.

Intermittent Streams and Permanent Streams. — While in the gully stage, the valley is likely to carry water only during and after rains.

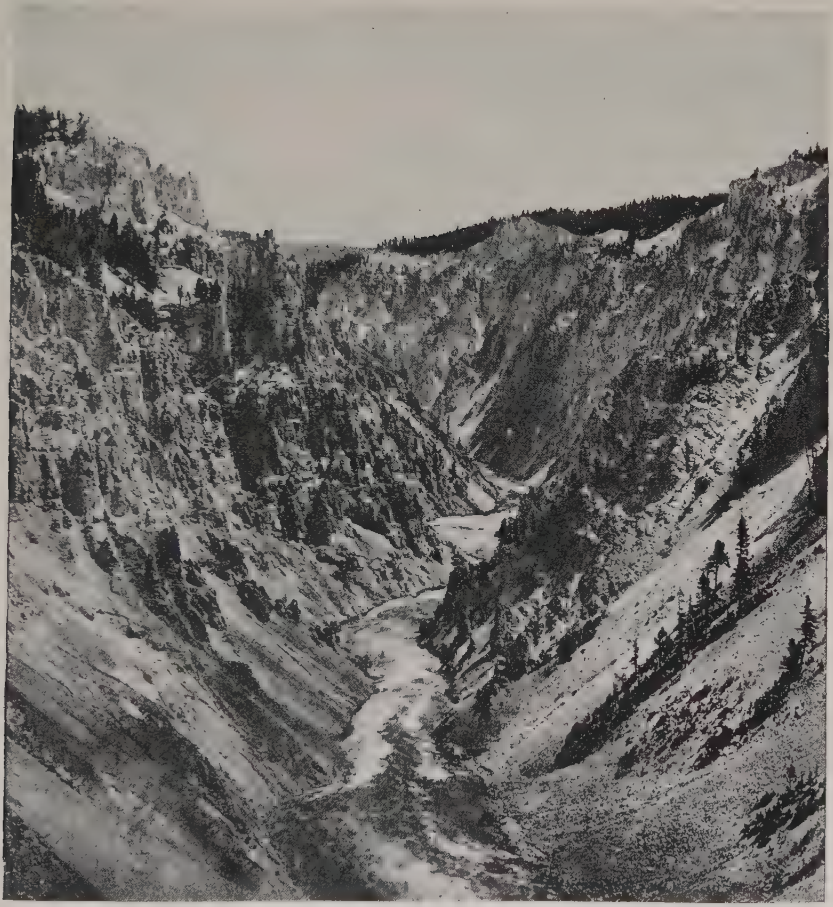


Fig. 28. — A valley in a youthful stage of its history. Yellowstone River. (U. S. Geol. Surv.)

Under these conditions the stream is said to be *intermittent*. As erosion deepens the gully, however, more and more of the rain water that enters the ground directly has opportunity to emerge again in the gully sides, and to contribute to the flow of water on the gully bottom long after the parent rains have ceased. Eventually the gully (perhaps a valley now) is excavated to and below the level at which all openings in the rock are permanently filled with ground water (Chapter VI). From this time on the water seeps steadily and uninterruptedly into the valley bottom, and the resulting stream is said to be *permanent.*

Baselevel. — When a stream reaches the sea its velocity is checked, its load is deposited, it can cut downward no further, and consequently it can erode no more. The level of the sea, projected inland as an imaginary plane below the surface of the land, is called *baselevel* because it is the ultimate base and goal of denudation by streams. It follows that a stream hastens its own downfall every time it removes a cubic yard of material from any part of its bed, because thus by just so much it lowers the gradient on which its velocity depends. In Fig. 23 the base line represents the baselevel toward which the gradient *ab* is steadily being lowered.

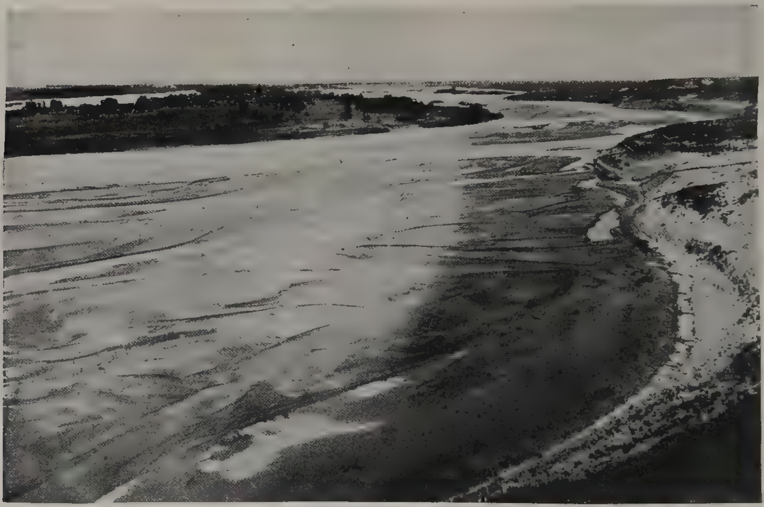


Fig. 29. — A heavily loaded river. Note the wide bed with many shallow interlacing channels and very broad valley. North Platte River above Gering, Neb. (U. S. Geol. Surv.)

Grade. — Since in those places where the gradient is lessened a stream tends to deposit, and erosion again sets in when the gradient increases, it follows that, as time goes on, a river proceeds to fill up the hollows and to cut away the projections in its bed and thus to establish a definite gradient. The gradient which the river seeks to establish is one at which, in each part of its course, the velocity is sufficient for the volume of water there present to transport its burden without erosion or deposition; the stream is then said to be *at grade.* This does not mean that the gradient is uniform from source to sea; it may be relatively much steeper in the upper course, where the load consists of coarse débris and the volume of water is small, than in the lower part where the slope is gentle but the volume of water is large and the load consists of fine sediment. A heavily loaded stream, like the Platte (Fig. 29), may become graded on a relatively steep slope, as compared with one, not fully loaded, which on such a slope would be ungraded and still degrading. The lower parts of great rivers such as the Mississippi become graded while, in their headwaters, cutting and deepening by erosion are still actively going on.

Thus in summary we may say that a stream is *at grade* when its transporting power and the load given it to carry are about equal. It is *aggrading* when the load it has to carry exceeds its ability to transport. It is *degrading* when its ability to do work is in excess of the material to be carried, and the excess of energy is employed in deepening its channel. When the stream has reached grade it is said to have reached its "profile of equilibrium" — the profile which permits a transporting power closely adjusted to the amount of waste to be carried.

Relation of Tributaries to Main Streams. — Examination of drainage systems shows that in a vast majority of cases the tributaries of a river enter it at grade, i.e., at the same elevation as the main stream. They are thus said to be *accordant.* The reason for this is that as the main stream cuts down its bed, the resulting increased gradient which is given the tributaries enables them to keep pace in downcutting in spite of the smaller volume of water they contain. But this may increase the ratio of the trenching of the lateral valleys over their widening to a greater degree than in the main trunk valley and hence they grow proportionately narrower and steeper walled. Examples are afforded by some of the tributaries of Colorado River. In their effort to keep accordant relations with the main stream they have cut narrow slot-like canyons. In some cases, however, small tributary streams, unable to keep up with a rapidly downcutting river, are obliged to cascade down the main valley walls.

Falls and Rapids. — Falls and rapids are common in the valleys of swift streams. Most of them are the result of the unequal erosion of rock masses composed of both hard and soft layers. This is magnificently illustrated in the great cataract at Niagara (Frontispiece). Niagara River, which drains the four upper Great Lakes, in its course of 36 miles from Lake Erie flows over a plateau which terminates near Lake Ontario in an escarpment more than 300 feet high. The plateau is capped by a resistant layer of limestone under which are soft, easily eroded shales. These relations are shown in Fig. 31. By the gradual disintegration and undermining of the softer underlying shale the harder limestone on top is left projecting as a lip over which the water falls (Fig. 31). From time to time this projecting rock, left unsupported and penetrated by joint cracks, also falls and is carried away. By this means the falls maintains itself and at the same time steadily moves upstream, leaving a deep gorge behind it. Originally the falls was situated at Lewiston near the mouth of the river, and falling over the escarpment had its full height at this point. It is now 7 miles above its original position (Fig. 30). Successive surveys made throughout a period of 50 years have shown that the falls is retreating at a rate that averages 5 feet per year.

Falls are caused in other ways as well; by glacial erosion, by the accidental damming of streams back of lava flows and landslides, and by uplift of the land relative to the sea. But whatever their cause, falls cannot indefinitely persist. They are doomed to disappear as the streams reduce their valleys to grade.

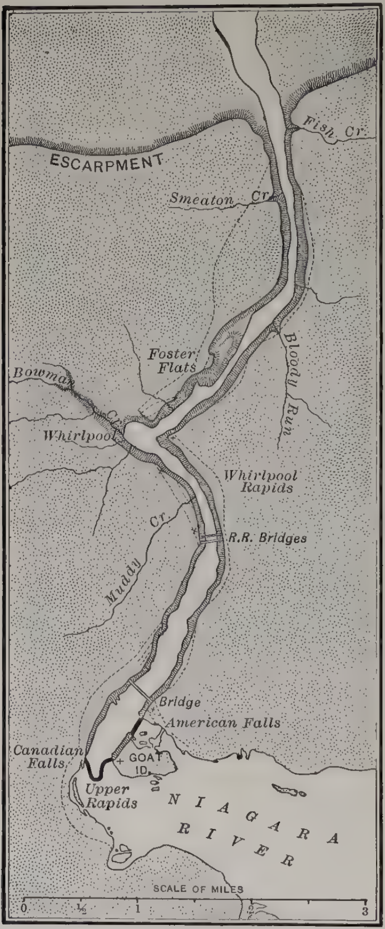

ESCARPMENT
Fish Cr.
Smeaton Cr.
Foster Flats
Bowman Cr.
Whirlpool
Bloody Run
Whirlpool Rapids
R.R. Bridges
Muddy Cr.
Bridge
American Falls
Canadian Falls
GOAT ID.
Upper Rapids
NIAGARA RIVER
SCALE OF MILES
0
½
1
2
3

Fig. 30. — Map of Niagara River and Falls. (After G. K. Gilbert.)

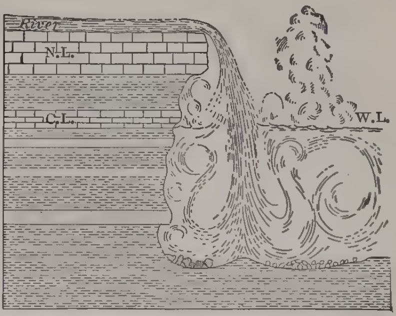

River
N.L.
C.L.
W.L.

Fig. 31. — Section showing rock layers and cause of falls at Niagara. (After G. K. Gilbert.) N. L., Niagara limestone with soft shale below; C. L., Clinton limestone with shales and sandstones below. 1 inch = 300 ft. W. L. = water level of pool.

Valley Widening. — Every stream valley is more or less crooked and the current, forced to follow a winding route, is thrown against the outside wall at every bend. While the stream is youthful and vigorous its tendency to cut laterally against the valley wall is strong; but at this stage of its development the stream is cutting down rapidly to establish grade, and therefore the stream bed does not remain long enough at any one level to permit lateral cutting to accomplish much. After grade is established, however, this situation is changed. Further deepening of the valley then proceeds very slowly; and as the current is deflected from side to side the valley walls begin to recede appreciably from the attack. As the bottom of the valley widens by this process the cross section of the valley loses the **V**-form typical of the youthful stage (Fig. 28) and takes on a **U**-shape characteristic of maturity. In time the valley floor becomes wide enough to accommodate a large flood plain, and the curves of the stream touch the valley bluffs only at isolated points (Fig. 32). However the curves shift their positions slowly, as explained in a later paragraph, and continue the cutting action against all parts of the valley sides. This process of valley widening by lateral cutting of the stream is called *lateral planation.* Its results appear first in the lower portion of a valley, where stream grade is established earliest. Gradually the widening progresses headward as the valley grows older.

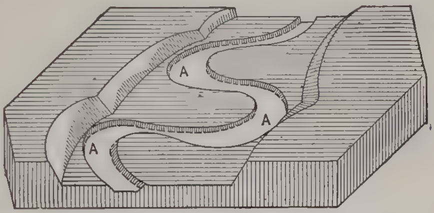


Fig. 32. — A river widening its valley by meandering and planation.

CONSTRUCTIVE WORK OF RIVERS: STREAM DEPOSITION

So far in the study of rivers we have considered the destructive, erosional work they perform — work done largely in their upper reaches. Some rivers have a swift course through elevated tracts of country to the sea, their work is cut short when they enter it, and they deposit their load at once; but many, and especially the larger rivers of the world, descend into wide lowlands, through which with steady current they wind to their journey's end. In these lowlands, and at the rivers' mouths, the work done is different from that in the upper reaches; it is largely constructive, rather than destructive, and consists mainly in the deposition and slow shifting of the burden assumed through erosion in the higher part of the course.

Flood Plains; Natural Levees. — As the gradient of a stream steadily lessens, and as the valley widens, the sediment dropped is spread continuously along the valley bottom, building up great flats. In the spring, melting snows and heavy rains pour a vast volume of water into the tributaries of the stream. This volume, concentrated in the main stream, raises the water level until it overtops the channel and floods the bordering flats. The water which has been moving swiftly through a deep channel, with friction at a minimum, thus suddenly forced out on to a shallow flat, is quickly and almost completely checked. Concentrated deposition therefore takes place along the immediate borders of the flooded channel, grading outward away from the stream into much thinner deposits. In consequence, low ridges are built up paralleling and confining the stream channel; and these remain after the flood has subsided and the stream has shrunk to its old dimensions. They are known as *natural levees* (Fig. 33). Beyond them the land is low and usually swampy. The whole broad flat, including both the natural levees and the flanking lower lands, is called the *flood plain.* The flood plain of the Mississippi covers an area of about 30,000 square miles, and a large portion of it consists of extensive swamps.

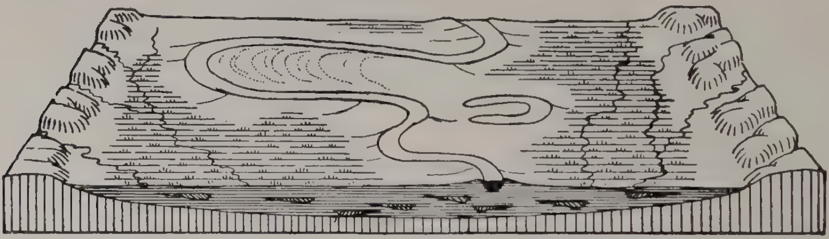


Fig. 33. — Flood plain with natural levees built by a large stream.

With the subsidence of the flood and consequent decrease in volume of water the stream not only stops scouring its channel; it actually silts up its channel with sediment. In this way the stream may gradually come to flow at a level higher than that of the flats beyond the levees, being restrained *only by the levees* from deserting its course and occupying the flats. This is the situation indicated in Fig. 33 and is moreover the condition in the lower Mississippi when in normal flood. This has become a serious matter from the point of view of floods, with the gradual settlement and agricultural development of the low, ill-drained areas back of the levees.

Meanders. — A stream of low gradient flowing through a flood plain is easily turned by obstacles that it lacks the power to remove. Any deviation from a straight course throws the current against one of the stream's banks. Some erosion of the bank occurs, and the current is thrown against the opposite bank (as in *A*, Figs. 34 and 35) where again erosion takes place. Symmetrical curves are thus formed, and a continuation of the process increases them. Meanwhile the current is slackened at *B* (Fig. 34), deposition occurs there, and the inside of the curve is built out as the outside is cut away. In this manner the arcs of the curves become more and more pronounced as *CC* and *DD*. Eventually a loop, as at *E*, is cut through, leaving an island in the stream, the main current takes the shortest route *FF*, and the entrances to the abandoned channel are quickly silted up, leaving a shallow crescentic lake. The symmetrical curves are called *meanders* (from the River Meander in Asia Minor) and the resulting lakes, *oxbow lakes*. Meanders are roughly proportional in size to the size of the stream with which they are associated. The arcs in small streams commonly have circumferences of only a few score feet, but some in the Mississippi are 20 miles or more in length.

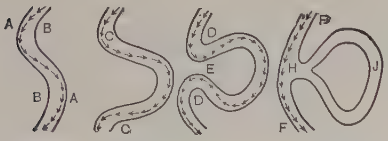


Fig. 34. — The formation of meanders and ox-bows.

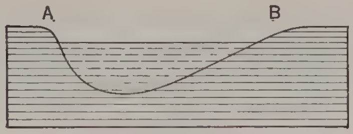


Fig. 35. — Section of a stream channel from *A* to *B* shows it to have the profile seen above, the deepest part lying close in to the bank at *A*. See Fig. 34, *A* and *B*.

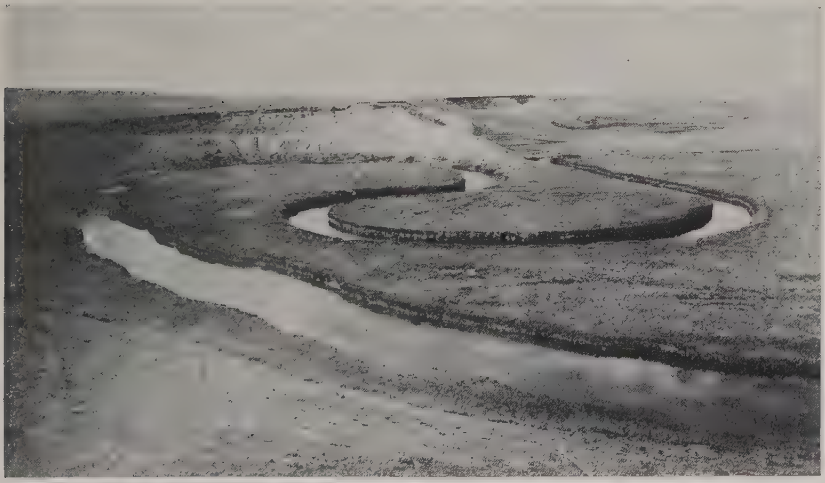


Fig. 36. — Meanders of a stream in nearly flat region. Trout Creek, Yellowstone Park. (U. S. Geol. Surv.)

Lateral cutting is necessarily more effective against those portions of the meander curves which face upstream than against those which face downstream; hence each meander loop tends to shift slowly downstream during its existence. Thus a constant succession of meanders moves almost imperceptibly mouthward past any given point in a stream valley. This steady shift downstream is called *sweep*.

Artificial Levees; Floods. — Before the settlement of the lower Mississippi Valley, the river was in the habit of overflowing its banks frequently, the floodwater reaching the sea through the low flats beyond the natural levees. As early as 100 years ago attempts were made to confine the flooded river to its channel, in order to reclaim for agricultural purposes the adjacent rich and fertile alluvial lowlands. These attempts took the form of artificial levees which were built upon the natural ones, thus deepening and heightening the channel at the same time. If a given amount of water is confined to a narrowed course and thus prevented from normal spreading, its channel must necessarily be deepened. A part of this added depth is attained locally by increased erosion of the bottom, but much of it is attained by rise of the water surface. Hence whenever a new levee is put in, all those already in existence must be raised. The first levee, built at New Orleans, was 4 feet high. Today the average height is more than 13 feet. In 1902 there were 1300 miles of levees along the Mississippi; in 1927 there were 2500. The river level is gradually rising higher above the adjacent basins, and when floods do occur, they are correspondingly more destructive. The building of levees is therefore not in itself an adequate method of preventing floods.

The scheme of building levees has been followed in many parts of the world. In the flood plain of the Po the river bed is above the housetops. Breaks in the levees on the Hoangho at various times have resulted in enormous loss of life. In the flood of 1887 the Hoangho, called " China's Sorrow," drowned considerably more than a million people.

Deltas. — When the current at the mouth of a stream is checked by a body of standing water such as a lake or the ocean, its load is promptly dropped on the bottom, building an embankment with a front which grows outward like a railroad or highway fill in process of construction across a valley. As the deposit is built up close to the water surface, the flood plain usually encroaches upon it from upstream and gradually covers it so that it is built above water in a crudely triangular shape with one apex pointing upstream. From this shape, resembling the Greek letter, the deposit derives its name of *delta*.

The bulk of the delta-forming material is dropped on the frontal slope of the growing embankment, forming thick *foreset beds* (Fig. 37). Some of the finest sediment however remains longer in suspension and is carried farther out and dropped as *bottomset beds* which thin outward away from the delta. Along the top, where erosion and deposition alternate as the stream current changes seasonally, thin horizontal *topset beds* are laid down. The thickness of the whole mass is in some cases very great. Borings put down in Venice, which is built on the delta of the Po, reached a depth of more than 500 feet without attaining the bottom of the delta beds.

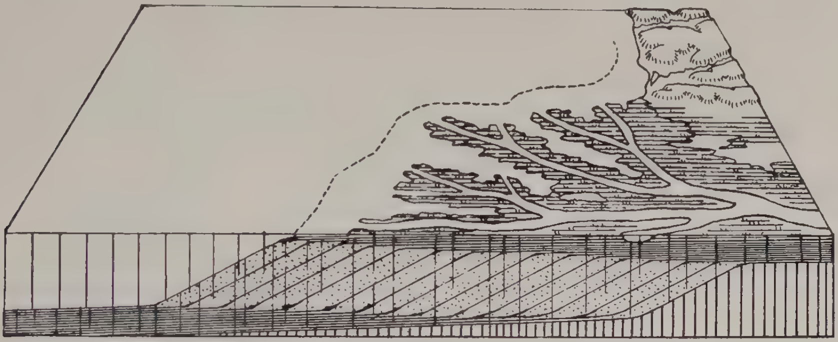


Fig. 37. — Ideal plan and section of a delta. Plan shows delta fingers and distributaries. Section shows thick, steeply dipping foreset beds overlain and underlain by thinner topset and bottomset beds, respectively. Some sinking of the growing delta with relation to sea level is indicated by the arrangement of topset and foreset beds.

Whenever the river in flood overtops the natural levees, it has opportunities to break through at certain points and to flow seaward through some new channel, leaving a diminished volume of water to escape through the old channel. Since this happens so frequently as to be the rule in the case of large rivers left to themselves, new outlets are broken through soon after they have been formed, and a branching system of distributaries grows up, giving shape to the delta (Fig. 39). Thus a long stream ending in a delta is like a rope frayed at both ends, the strands at the upper end being represented by the tributaries and the shorter ones at the lower end by the distributaries. The branching system extending seaward is the skeleton of the growing delta; between the long arms lie shallow basins which gradually fill with sediment during floods and thus become low land.

The shifting of the main channel through the development of new distributaries is strikingly illustrated by the case of the Hoangho, which for approximately 700 years prior to 1852 had discharged eastward into the Yellow Sea. In 1852 it broke its banks at a point more than 300

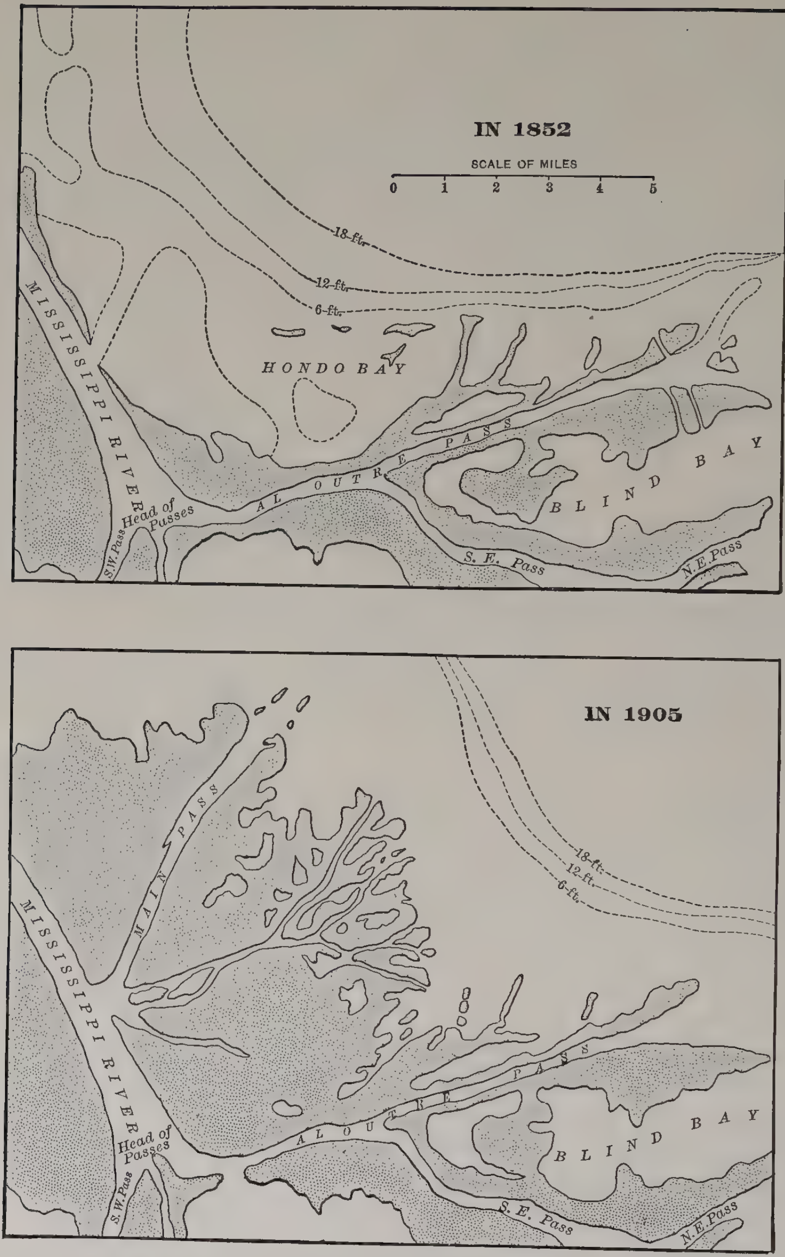

IN 1852
SCALE OF MILES
0 1 2 3 4 5
18-ft.
12-ft.
6-ft.
HONDO BAY
MISSISSIPPI RIVER
Head of Passes
S.W. Pass
A L O U T R E P A S S
BLIND BAY
S. E. Pass
N.E. Pass
IN 1905
18-ft.
12-ft.
6-ft.
MAIN PASS
MISSISSIPPI RIVER
Head of Passes
S.W. Pass
A L O U T R E P A S S
BLIND BAY
S. E. Pass
N.E. Pass

Fig. 38. — Growth of part of the Mississippi delta during 50 years. (After G. R. Putnam.)

miles above its mouth, formed a new channel northeastward across the great alluvial flats of the province of Shantung, and finally emptied into the gulf of Chihli, almost 300 miles north of its old mouth. The Hoangho has occupied this new course with minor distributaries since 1852, and the old channel has largely dried up.

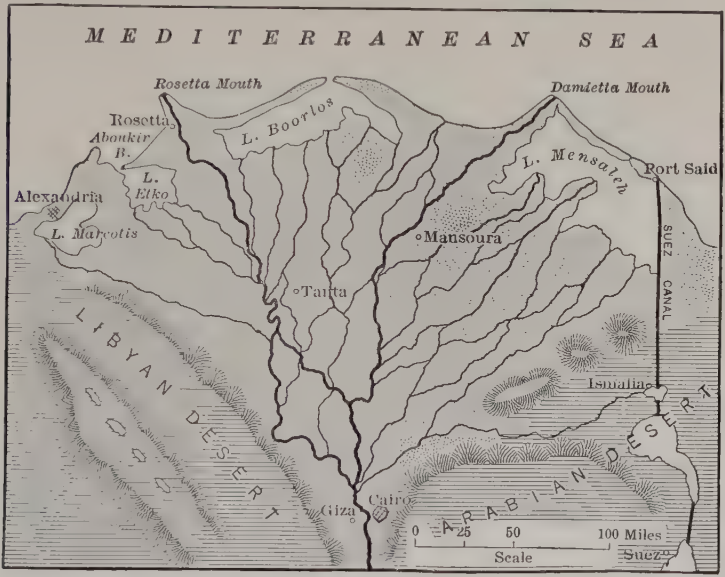

MEDITERRANEAN SEA
Rosetta Mouth
Damietta Mouth
Rosetta
Aboukir B.
L. Boorlos
L. Mensaleh
Port Said
Alexandria
L. Etko
L. Mareotis
Mansoura
SUEZ CANAL
Tanta
LIBYAN DESERT
Ismailia
ARABIAN DESERT
Giza
Cairo
0 25 50 100 Miles
Scale
Suez

Fig. 39. — Delta of the Nile, showing its form and distributaries.

Stream Terraces. — Many stream valleys, particularly in their middle reaches, are bordered by bench-like flats the tops of which are higher than even the flood stages of the present streams. In many cases there are several of them in series, rising away from the river like two long flights of steps facing each other. In some valleys they are continuous for long distances. Such flats, high and dry, are called *stream terraces*. Upon examination of these forms it soon appears that some are made of rock (*rock terraces*) whereas others apparently contain no bedrock but consist entirely of unconsolidated sands, gravels, and clays deposited by the streams when they were flowing at higher levels (*flood-plain terraces*).

Rock Terraces. — These are of two kinds. (1) Rock terraces found usually in dry regions where valleys have been cut down through alternating layers of weak and resistant rocks. The resistant layers, less easily eroded as the stream cuts down, are left standing out in relief, while the weak layers are etched back (Fig. 40). The terraces in the

Grand Canyon are an excellent example. (2) Rock terraces caused by successive uplifts of the land. These are mentioned here only for the sake of completeness and are discussed in connection with changes of level (Chapter XII).

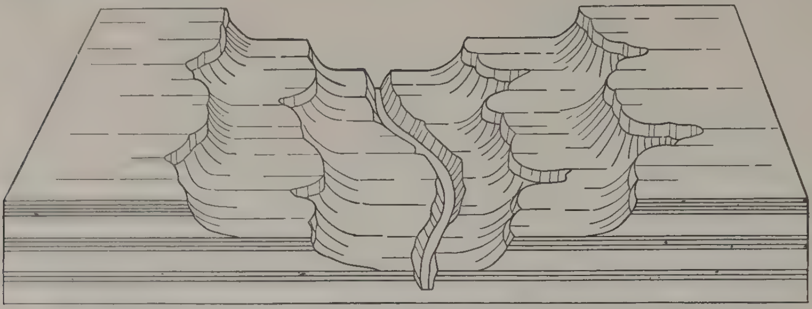


Fig. 40. — Rock terraces caused by differential erosion of a series of strata of unequal resistance.

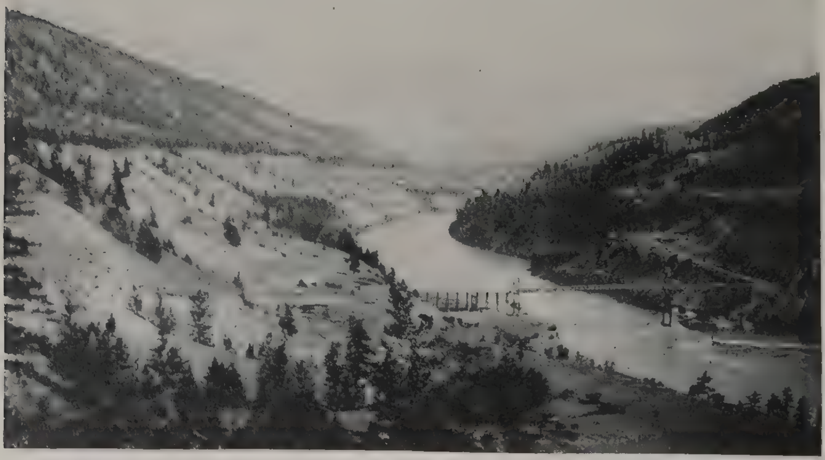


Fig. 41. — Terraces on the middle Fraser River, B. C. (F. F. Osborne.)

Flood-plain Terraces. — Flood-plain terraces, the most common type, are formed by a stream which has begun to degrade following a long period of aggradation in which a flood plain was built up. It is obvious that any increase in the gradient or volume of an aggrading stream or decrease in its load would cause it to begin to degrade and hence to leave terraces (Figs. 41 and 42). A sudden upward warping of the land while streams are flowing over it, increasing their gradient, is regarded as a common cause. Another cause not rare in glaciated regions is the melting away of glaciers and the consequent great diminution of the load of the streams that drained them. Many terraces also are *rock-defended*. They are developed in this way: A stream of low gradient which has been slowly excavating the deposits in the bottom of its valley, and is meandering from side to side across its valley floor, encounters at numerous points the bedrock of the valley wall. Deflected by this unexpected obstacle, it swings away and thus delays its undercutting of the remnant of the old flood plain above the protective rock outcrop, which remains for a time as a terrace, gradually increasing in height as the stream continues to cut downward (Fig. 43).

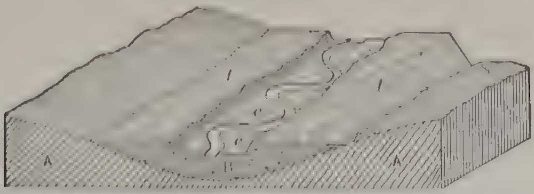


Fig. 42. — Formation of flood plain terraces. *AA*, section of river-cut valley; *B*, alluvial deposits of river; *tt*, former flood plain, now forming terraces; *c*, new flood plain.

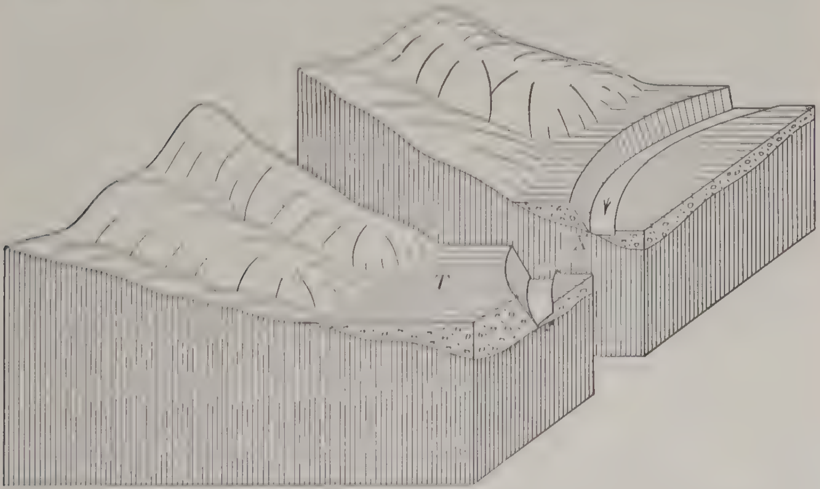


Fig. 43. — Old flood plain "defended" against undercutting by a rock outcrop at *X*, thus gradually forming the terrace *T*.

Alluvial Cones and Fans. — When a swift tributary stream enters a wide and nearly level valley, the abrupt change in its gradient may cause it to deposit the greater part of its load on the valley floor. In this way a low semi-conical elevation is formed, radiating out from the mouth of the tributary (Fig. 44). Such forms are generally known as *alluvial fans*, but if steep they are sometimes called *alluvial cones*. In one sense they may be regarded as deltas formed on land, but they differ from deltas in having a sloping rather than a flat top, since their upbuilding is not controlled by the surface of a body of standing water. An alluvial fan acquires its shape because of uniform distribution of débris over its surface by the parent stream. The stream repeatedly silts up its channel, overflows, and forms new distributaries. Thus when one part of the fan is built up, the stream shifts to another course (which of necessity is temporarily lower) and builds that up. In this way the entire surface eventually is covered by the stream.

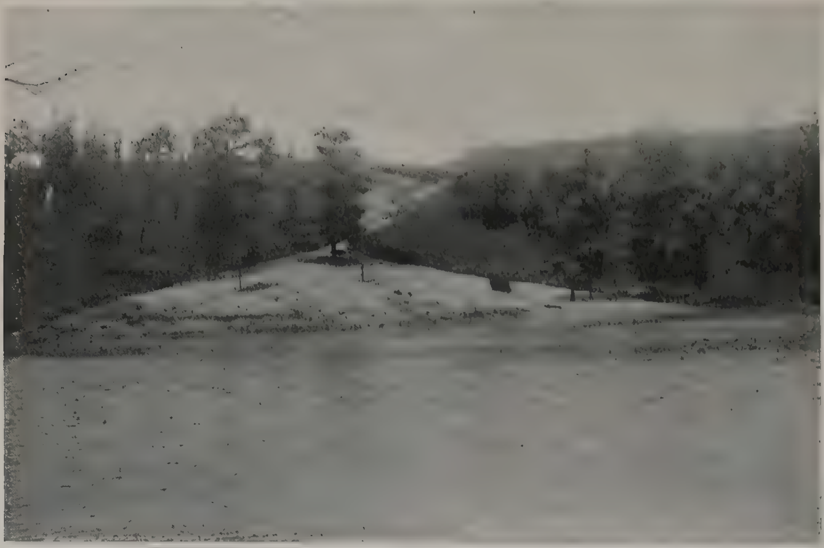


Fig. 44. — Alluvial cone, made by a tributary to a larger stream. Stoughton, Wis. (U. S. Geol. Surv.)

Alluvial fans are not common in humid regions. Where present at all they are small, and are usually made of soft material easy to erode, such for example as glacial deposits. In many dry regions, on the other hand, deposits of this type are characteristically large, and they are conspicuous features of arid landscapes.

THE CYCLE OF STREAM EROSION

Since streams and their tributaries steadily develop and enlarge their valleys, and since cutting by the whole system is limited downward by a *baselevel*, what will be the final result of long-continued erosion by streams on a given land mass? Obviously there can be but one answer: the wasting away of the land to a low, gently sloping surface from which the water must drain sluggishly to the sea. This process, the complete denudation of a land mass, of course involves an enormous amount of time, but the final result is inevitable. The series of changes involved in the complete reduction of a region to baselevel constitutes a *cycle of erosion.* The time required necessarily varies with varying circumstances such as initial elevation above the sea, resistance to erosion of the underlying rock, and amount of rainfall and run-off.

THE CYCLE OF STREAM EROSION IN A HUMID REGION

Although the process governed by the cycle is continuous, it is divided for convenience into stages, as follows: *initial stage, youth, maturity, old age.* It must be remembered that each of these stages grades into the next, and that all are parts of one unbroken chain of events. In connection with the following account of a typical case, reference should be constantly made to Figs. 46-51.

Initial Stage; Consequent Streams. — An upwarped area of sea bottom, bearing on its surface initial irregularities, appears above the sea. For the sake of simplicity let the material composing the mass be broadly homogeneous, with only local variations in its resistance to erosion. Gullies develop (Figs. 45 and 46) and grow mouthward and headward under the control of gravity and the initial slopes. Streams whose development is thus controlled by original surface irregularities are called *consequent streams.* Adjacent gullies grow into one another on favorable slopes, forming connected chains. They gradually become ravines and then valleys. In the early stages of the cycle, the gradient is so steep that downcutting of the valley bottoms is dominant over slope wash on the valley sides; hence the valleys and gullies are steep-sided and sharply V-shaped. Any irregularity of slope or material in the side of a gully is likely to concentrate run-off and thus to develop a tributary. Since no gully is uniform in these respects, tributaries rapidly develop, all lengthening themselves headward from the parent gully. When the initial gullies have developed initial tributaries, they break up the continuity of the initial slopes, and the land area passes from the *initial stage* into *youth.*

Youthful Stage. — The land area is now drained by an integrated drainage system consisting of main streams developed from the pre-existing initial slopes. These slopes were so arranged that more concentration of drainage took place along the lines of the infant streams *B* and *C* (Fig. 47) than along streams *A* and *D.* The chain of cause and effect was thus set up most rapidly and thoroughly along *B* and *C* and these streams sent out more tributaries and grew headward more swiftly than did *A* and *D*. Briefly stated, a struggle for existence takes place among adjacent streams in the competition to grow big and therefore to absorb larger drainage areas. In this case, the streams (*B* and *C*) favored by the initial slopes win out and maintain their lead over their less favored and therefore weaker neighbors (*A* and *D*).

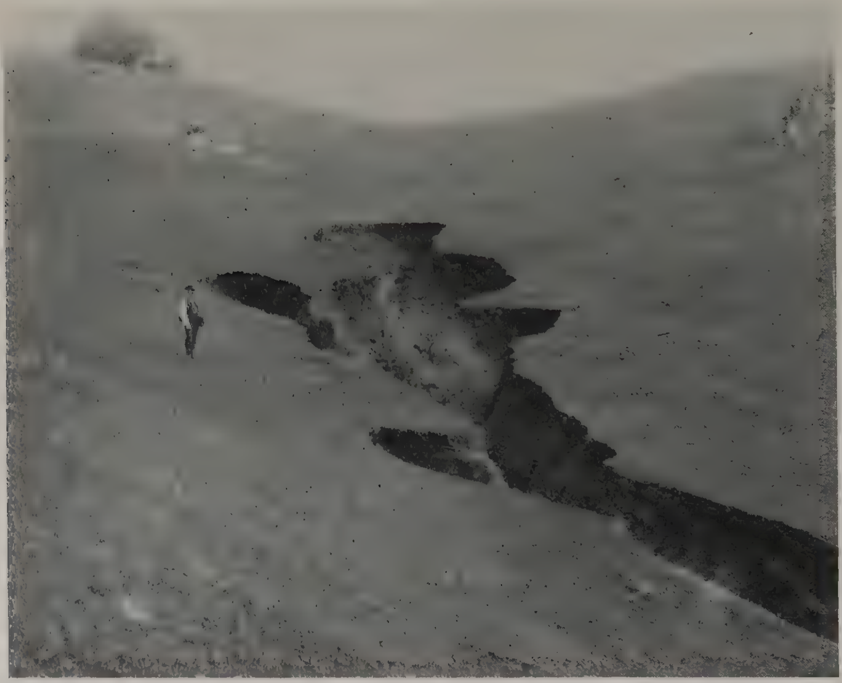
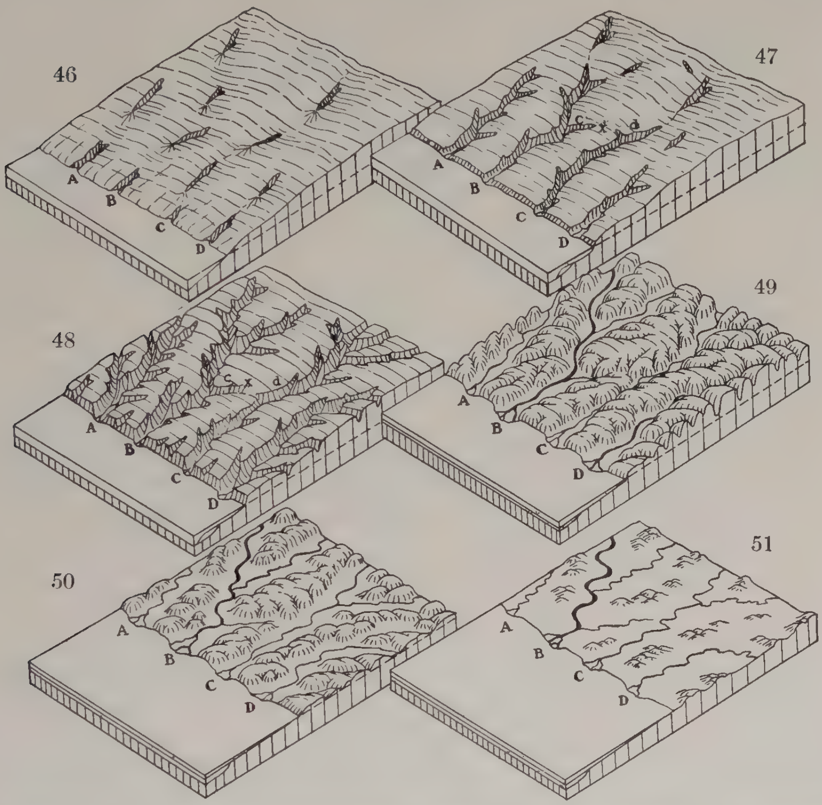


Fig. 45. — A newly-formed gully, starting the dissection of a smooth surface. Note the embryonic tributaries, which will grow longer by headward erosion as the main gully extends farther up the slope by the same process. (U. S. Geol. Surv.)

Divides. — The area of higher land between two valleys is called a *divide*. The divide between two parallel gullies of unequal size may be completely destroyed by the growth of the larger gully (Fig. 52). The divide between the heads of two streams flowing in opposite directions is easily shifted laterally. Such a situation is developing at x (Fig. 47). Two streams, c and d, tributary respectively to *B* and *C*, have worked headward toward each other, thus narrowing the broad divide which formerly separated them. Stream c has a shorter journey to the sea than has stream d; hence the gradient of c is steeper; hence its power to corrade and to transport is greater; hence it eats both downward and headward more rapidly than does *d*. This inequality in rate of erosion results not only in lowering the divide *x* but in shifting it away from *c* toward *d* (Fig. 53). When erosion by *c* is equaled by erosion by *d* the divide ceases to move laterally and further erosion can only lower it. The divide is then said to be *permanent.*



Figs. 46–51. — Ideal cycle of stream erosion under a humid climate and in homogeneous rocks.

Fig. 46. — Initial stage, showing gullies developing wherever the run-off is concentrated.

Fig. 47. — Early youth, showing integration of main drainage lines and growth of the stronger at the expense of the weaker streams.

Fig. 48. — Later youth, showing the reduction of the initial surface to irregular flat-topped ridges.

Fig. 49. — Early maturity, showing dissection of divides into flowing slopes and development of alluvial flats.

Fig. 50. — Later maturity, showing decrease in relief, lowering of slopes and widening of valleys.

Fig. 51. — Old age, showing development of peneplain with monadnocks.

The streams in the youthful stage have gradients as steep as the height of the land above sea level permits; and since they are energetically cutting into their valley bottoms, the valleys are V-shaped in cross section, with steep sides whose slope depends on the resistance of the rock composing them (Fig. 48). The valley courses are crooked with irregular bends, all of them determined by the initial irregularities of the land surface. Tributaries develop rapidly, their valleys working headward from the main streams like branches growing from the trunks of trees. In fact, so closely does the pattern of a stream system under these conditions resemble a tree with its branches and twigs, that it is called a *dentritic* pattern. Each tributary enters its main valley at a level with the main stream. Downcutting by the tributaries keeps pace with downcutting by the main streams, and thus the whole system is delicately balanced and adjusted throughout its extent. As the countless tributaries continue to dissect the initial surface, the broad initial divides contract into narrow and irregular ridges. The time of youth is the time of scenic grandeur in a landscape. Deep gash-like valleys (Fig. 28) and canyons with foaming rapids, precipitous cliffs, and high ridges are characteristic. But the very force that sculptured these forms will inevitably destroy them.

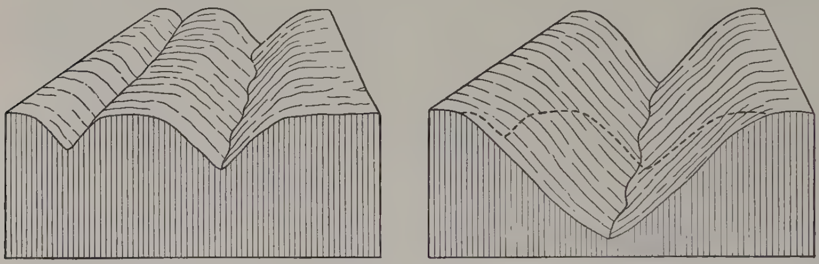


Fig. 52. — Lateral seizure of a small gully by a large one.

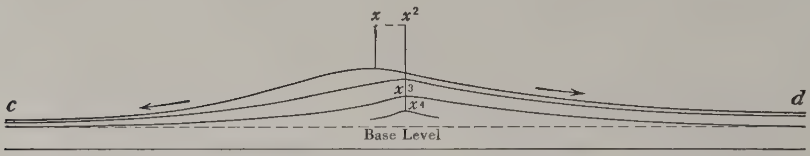


Fig. 53. — Evolution of a shifting divide into a permanent divide. Greater erosion by stream *c* than by stream *d* results in lateral shifting of the divide through the distance xx^2 as well as in lowering it. At this stage, erosion has become equal on both sides of the divide so that further downcutting results merely in lowering the (now permanent) divide to x^3, x^4, etc.

It is during the period of youth that landslides are common in regions where steep slopes and suitable climatic conditions are found. The sliding of great masses of rock material from the steep sides of mountains and canyons hastens materially the destruction of the land.

Mature Stage. — When the tributaries have worked headward and widened so far that they have destroyed all of the initial surface by converting it into slopes, the landscape takes on a wholly new aspect and the region is said to be *mature* (Fig. 49). The intricate network of drainage is complete. The main streams have cut downward far enough to decrease their own gradients appreciably. As decreasing gradient progressively decreases each stream's downcutting power, and thus causes it to linger at each successive level, the force of the current deflected from side to side of its valley begins to cut effectively and each valley is thereby widened. In this way valley widening increases proportionately as valley deepening decreases. Long stretches of the main streams are at grade as they wander over their newly developed valley flats, while the upland (initial) surface, so prominent during youth, has melted into slopes. Since downcutting has been greatly checked, the mantle of loose material on the valley sides is not swept away as rapidly as it is formed by the agents of weathering. It therefore accumulates, and moves slowly down the slopes under the influence of gravity (locally aided by slope wash, ground water, and frost action). In this way hollows in the slopes are filled in, irregularities are smoothed out, and the profiles of the valley sides are converted into smooth flowing curves (Fig. 54). It is for this reason that the period of maturity is the time of restful beauty in a landscape. The rugged splendor of youth has been modeled into sweeping mellow curves.

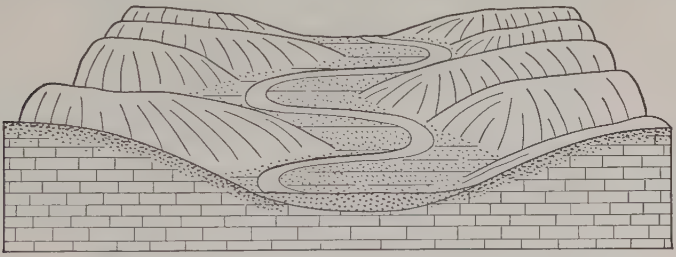


Fig. 54. — The smooth curving profiles of maturity, showing their relation to accumulated mantle rock which masks the horizontal strata. Compare the valley shown in Fig. 40, in which the horizontal strata appear unmasked because of continued rapid downcutting by the stream.

Old Age Stage. — Because of the low and ever-decreasing gradients of the main streams, the heights are now wasted by the steeper tributaries much more rapidly than the larger valleys can be cut down.

The result is a gradual decrease in relief as the divides are worn down to lower and lower levels (Fig. 50). Lateral planation by the main streams widens the valleys and thus helps to cut away the adjacent higher land. The streams are sluggish, meandering widely. Natural levees bordered by swamps are developed in their lower courses. As the bottoms of the main valleys slowly approach baselevel, corrasion by the main streams gradually ceases, and only the divides, where the gradients are still appreciable, are notably lowered. In this way the hills of maturity melt down into low elevations in old age, shedding the run-off feebly in sluggish streams. Erosion takes place more and more slowly. The last few feet of vertical cutting might require a longer time than the entire amount of preceding excavation. The resulting surface of low relief, very gently undulatory, is called a *peneplain* ("almost a plain," Fig. 51). The highlands have been brought low; the rocks that compose them have been carried bit by bit to the sea and have been there deposited in beds as sediment. Only a few residuals of the former high land remain. Here and there isolated hills (Fig. 51) rise above the general surface, like islands above a sea. They are composed either of very resistant rock or of masses so far from the main streams that they only stubbornly allow themselves to be graded down to the general level of their surroundings. Such island masses are called *monadnocks* after Mount Monadnock in New Hampshire which rises in this manner above the level of the surrounding country.

It must be clearly understood that the topographic terms *youth*, *maturity*, and *old age* do not refer to periods of years, or to any absolute age. They denote merely stages, defined by the amount of work done in proportion to the total amount of work involved in the cycle. Thus a region of very soft rocks might reach old age while an area of resistant rocks was still in youth as far as the amount of erosion accomplished is concerned. It follows that an extensive valley system might exist in various topographic stages in different localities, depending on supply of water and the varying nature of the underlying rocks. As a matter of fact this is true of the valleys of most large rivers.

Effect of Vegetation on Erosion. — In humid regions the surface generally is covered with an almost continuous blanket of sod, supplemented locally by brush and forest. This mat of vegetation occupies uplands, slopes, and valley bottoms to the very edges of the streams. Through the action of frost and through repeated saturation with ground water, aided by gravity, the surface soil creeps down the slopes, carrying the vegetation with it. The movement is so slow and imperceptible that the mat of sod is rarely breached. Fresh gullying is hindered for several reasons: (1) Because the mass of roots distributed through the soil, together with the mat of organic matter on the surface, holds the soil firmly together and enables it to resist the pressure of the moving water. (2) Because the mat of vegetation acting like a sponge absorbs the water and permits it to drain off so slowly that the erosive effect of sudden rushes of water after storms is prevented. (3) Likewise in springtime the rapid melting of the snow is hindered by forest shade.

If the forest cover of a country is removed, erosion proceeds rapidly (Fig. 55), and in a variety of ways great damage may be done. The regulative action of the forests on erosion and the flow of rivers is a matter of great importance, not only from the geologic standpoint, but as vitally affecting civilization. In some countries, of which parts of northern China and Spain might be selected as examples, the improvident removal of the entire forest cover has reduced large areas, through displacement and loss of arable soil by erosion, to sterile wastes, subjected alternately to hot and baking droughts and sudden disastrous floods. Destruction of the forests by fire may have a similar effect. Once destroyed, and the soil washed out, they may be restored only with great difficulty after long periods of time. Considerable areas in the southern United States have been much impoverished in this way.

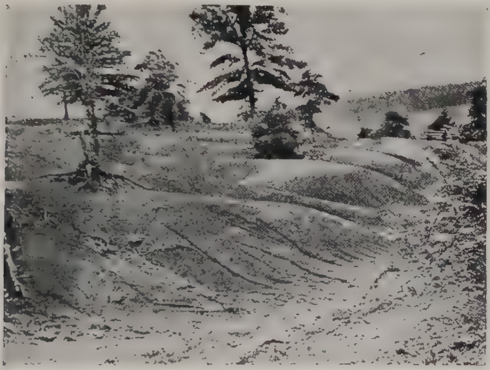
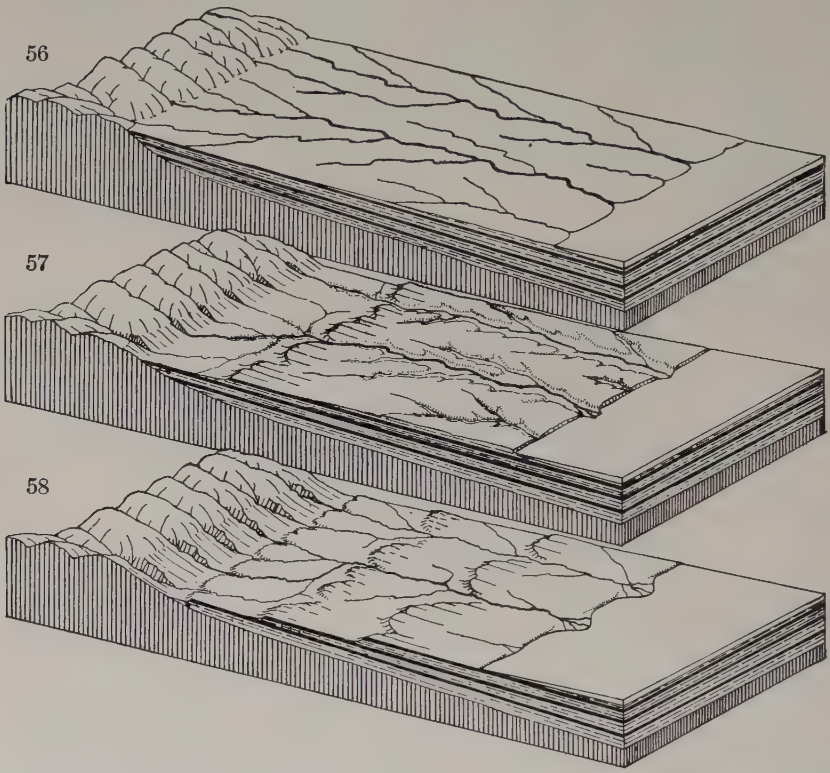


Fig. 55. — After the removal of the forest cover the soil has been carried away so rapidly that the remaining trees have their roots exposed by the lowering of the surface. Southern Appalachians. (U. S. Forest Service.)

Erosion on Tilted Strata. — In tracing the erosion cycle thus far, ideally simple conditions have been assumed. If the bedrock is uniform in resistance, or if it consists of horizontal sedimentary strata, erosion proceeds more or less evenly over an entire region and drainage develops with a normal dendritic pattern. Important modifications result, however, if the bedrock varies in resistance from place to place, and especially if there are tilted layers alternately resistant and weak.

This is the condition on a typical coastal plain formed by recent uplift of a shallow sea floor (Fig. 56). When the sea has retreated from the area, the result is a nearly even surface on gently-dipping sedimentary rocks. The plain is not unbroken, however, because the streams of the oldland must have extended themselves mouthward as the sea began to retreat, in order to keep pace with the retreating shoreline, cutting at the same time an ever lengthening trench in the newly exposed coastal-plain rocks. Such streams are known as *extended streams.* They are of course consequent upon the uplifted surface as well (page 65); hence they are called *extended consequents* (Fig. 56).



Figs. 56–58. — Development of relief on a newly uplifted coastal plain.

Fig. 56. — Newly uplifted coastal plain composed of gently dipping beds of unequal resistance, and drained by extended consequent streams.

Fig. 57. — Etching out of a cuesta from the topmost resistant bed by the development of subsequent streams.

Fig. 58. — Evolution into a belted coastal plain by the development of additional cuestas as lower strata are exposed.

As rain falls on the exposed coastal plain, tributaries develop, working headward from the trenches of the extended consequents. They develop most readily along zones of weakness in the underlying rocks. Such a zone is found in the contact of the recent shore deposits upon the much older, harder rocks of the oldland. Along this former land margin, then, the tributaries rapidly develop, working headward in a direction at right angles to that of the main streams. Such tributaries are called *subsequents;* all streams that work headward in this way along weak rock belts are so called. The parallelism of the weak belts in this case causes the subsequents to enter the main streams at right angles. The rectilinear drainage pattern thus developed is called a *trellis* pattern. (Compare the *dendritic* pattern developed in homogeneous rocks.) These subsequent streams cut the seaward, soft-rock sides of their valleys much more rapidly than the oldland sides where the rock is hard. Hence their channels constantly tend to shift seaward as by erosion they etch away the edge of the uppermost coastal-plain stratum. The product of this etching is a strongly asymmetrical low ridge, with a steep escarpment facing landward and a very gentle slope facing seaward formed by the surface of the gently dipping coastal-plain deposits (Fig. 57). Such a ridge is known as a *cuesta* (the Spanish word for this type of land form).

Coastal plains are favorable situations for the development of cuestas and trellis drainage patterns; but these land forms develop wherever alternately resistant and weak sedimentary formations are gently inclined. An excellent example is the great Allegheny Cuesta in central and western New York.

Belted Coastal Plains. — Wherever the extended consequents discover a resistant layer overlying a weaker layer in the sedimentary series, they send out subsequents which cut into the underlying soft rock, undermine the hard cap, and etch it slowly down its dip toward the sea. In this way a whole succession of cuestas may be developed, separated by the troughs of subsequent streams. Coastal plains bearing two or more cuestas are called *belted coastal plains* (Fig. 58).

The escarpment of a cuesta increases in height until the main consequents have become graded. At this stage downcutting by the tributaries has virtually ceased, and since they can thus cut only laterally, the cuesta escarpments are gradually lowered by being etched seaward, until they disappear. The result is a peneplain.

The coastal plain of New Jersey is belted, one of the belts being formed by the Navesink Highlands. Much of the country between Paris and the eastern frontier of France consists of a series of cuestas with east-facing escarpments. This arrangement was of the highest importance to the Allied armies on the western front during the Great War, since it presented a series of difficult natural obstacles to the German advance.

If the rocks have a very strong tilt, as in some mountain zones, other peculiar features are developed during the cycle of erosion. Discussion of these more complex land forms is postponed until the folding of rocks and the development of mountains have been considered.

The Cycle of Stream Erosion in a Semiarid Region

Vegetation and Erosion in Semiarid Regions. — The importance of the vegetation common to humid regions has been outlined in the preceding paragraphs. Let us turn now to drier regions where vegetation is less abundant. In semiarid regions (regions where the annual rainfall is roughly 10 to 20 inches), such as in much of the Great Plains region east of the Rocky Mountains, conditions somewhat resemble those in the deforested areas of the more humid country farther east. Trees are rare, the sod mat is present but not generally strong, and the soil is loose, dry, and porous. Rainfall, moreover, is likely to occur in sudden bursts. The result is rapid run-off and hence rapid erosion. Gullying occurs wherever the soil is laid bare, as on cattle trails and in wheel ruts. The streams are subject to sudden and heavy floods, their waters are very muddy, and in months of little rainfall they are commonly low or even dry. Because of the low rate of weathering of their side walls, coupled with rapid downcutting by their streams, the valleys are steep-sided, and many are flat-bottomed because of excessive deposition.

Effect of Resistant Rocks; Canyons and Gorges. — During the youthful stage of the cycle in humid regions, the streams cut down much faster than their valleys are widened by weathering and slope wash; and so if the initial elevation of the land is great, canyons and gorges (deep narrow valleys) may develop. But it is equally true that in maturity these must melt into wide, open valleys as downcutting decreases. In semiarid regions, however, the rate of valley widening by weathering is so slow that canyons and gorges, once they are developed, persist for a longer time. The grandest example is the Grand Canyon in Arizona, one of the most impressive wonders of the world. It is more than 200 miles long, 10 miles wide at the top, and from 3000 to 6000 feet deep. In general its cross section shows a broader upper canyon within which lies a deeper inner gorge (Fig. 59). It is cut in nearly horizontal beds of rock of varying hardness. These rest on crystalline rocks of almost uniform resistance, which in one stretch have themselves been trenched to a depth of 2000 feet in the inner gorge. The more resistant rock layers form gaunt cliffs whose talus slopes partly cover the softer beds, and whose outcrops form broad terraces and platforms. These effects, and the irregular cutting, carving, and recessing of the canyon walls through ravines and side valleys, have given rise to enormous and striking architectural forms (Figs. 2 and 249).

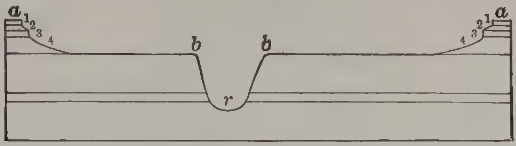


Fig. 59. — Ideal section across the Grand Canyon. (After Dutton.) *aa*, outer canyon walls; *bb*, inner gorge; 1 and 3, hard resistant beds; 2 and 4, soft beds.

Effect of Weak Rocks; Badlands. — When weak materials such as clays and shales are laid open to rain wash, they are rapidly carved into gullies. Striking examples of such erosion are to be seen along the rivers that drain the Great Plains region. These rivers, such as the Missouri and its tributaries, the Cheyenne and the Platte, in places run in valleys sunk a considerable distance below the general level of the country. The rock that forms the sides of the valleys consists for the most part of very soft, barely consolidated clays and sands, easily cut by rain wash and gullying. The result is that on either side of the stream from the bottom-land by the river to the bench-land forming the plain, there lies a gradually rising belt of country dissected in the

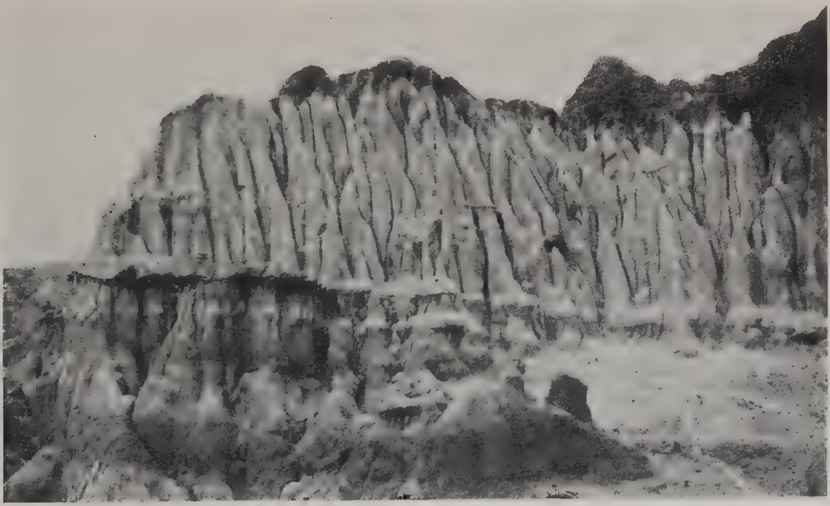


Fig. 60 — Effect of rain wash in beds of clay. Hell's-Half-Acre, Powder River, Wyoming. (U. S. Geol. Surv.)

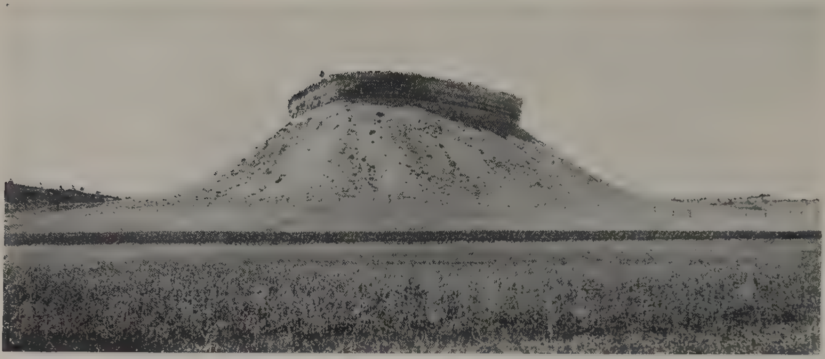


Fig. 61. — A butte near Billings, Mont. The capping is of sandstone, which resists erosion more successfully than the shale beneath it. (U. S. Geol. Surv.)

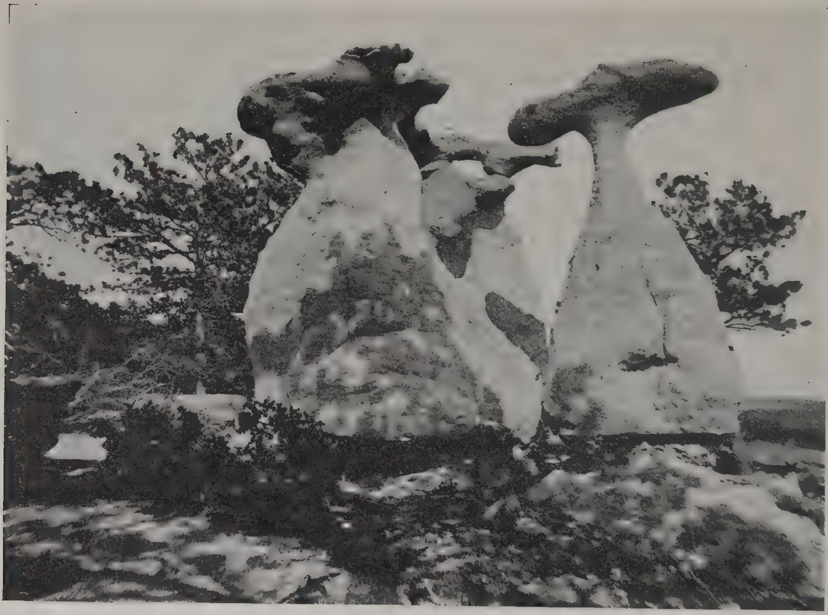


Fig. 62. — "Hoodoos" in Monument Park, Colo. Hard masses of ironstone in beds of soft sandstone have shielded the rock below them from erosion, and so pillars have been formed. (U. S. Geol. Surv.)

most intricate fashion by systems of gullies, gulches, and ravines, with spurs, knobs, and sharp ridges separating them. Such tracts of country are known as *badlands*, because of the difficulty experienced in traversing them (Fig. 60).

Sculptured Forms: Buttes and Mesas; Hoodoos. — In semiarid regions horizontal layers of hard rock resting on softer material do not weather to form rounded hills with gentle slopes. The edge of a resistant layer retreats as the weaker rock beneath weathers away, but it maintains a vertical cliff. Large isolated masses capped and thus protected by hard layers are known in the western United States as *buttes* (Fig. 61). Some are of mountainous size. Very broad flat-topped features, plateau-like in form, are termed *mesas* (Spanish "tables"). Many mesas are capped by lava. These features testify to the great amount of material carried away by erosion from around them.

Jointed rocks are eroded to form pillars, some of which have fantastic forms and are known as *hoodoos* (Fig. 62).

Summary. — The influences wrought by slight rainfall and consequent scanty vegetation impose themselves on the process of erosion throughout the semiarid cycle. The drainage pattern remains the same as in a humid region save that the tributaries are scattered instead of numerous. Alluvial fans play a much more prominent part in the semiarid than in the humid landscape, but following their extensive development in the youthful stage they are gradually cut away as the land is lowered, and the resulting surface is a peneplain.

THE CYCLE OF STREAM EROSION IN AN ARID REGION

Some authors classify arid regions roughly as regions that receive less than 10 inches of rainfall annually; others classify them as areas in which the drainage does not reach the sea. Chief among areas of *interior drainage* are the Sahara, the Libyan Desert, and the Kalahari in Africa, parts of the great Basin and Range region between the Wasatch Mountains and the Sierra Nevada in the western United States, the desert of western Australia, certain basins high in the Andes, and wide areas in central and western Asia. All are alike in that the streams which drain them lose themselves in the interior. But they differ in many respects. In some, streams are active agents of erosion and deposition, while in those that receive the least rainfall, the wind seems to be the chief dynamic agent. The latter have not been thoroughly studied. Because of this, and because no such deserts exist in North America, the discussion which follows is based upon the deserts of the United States.

Controlling Principles of Denudation in Arid Regions

Several dynamic processes are important in arid regions. The process of mechanical weathering is universal and is not further discussed. The other processes are as follows:

(*A*) **Stream Erosion and Deposition.** — Rainfall over the desert ranges sends down torrents of water through the dry gullies and gulches and out on to the plains below. The excessive load of loose weathered material acquired in the steep gulches is rapidly deposited in broad alluvial fans as the stream velocities are abruptly checked and as the streams lose volume through evaporation and sinking. Fans at the mouths of adjacent valleys coalesce and in time build up broad apron-like *piedmont plains.* The alluvial fan, comparatively rare in humid regions and common in semiarid regions, is the universal unit of stream deposition in regions of aridity.

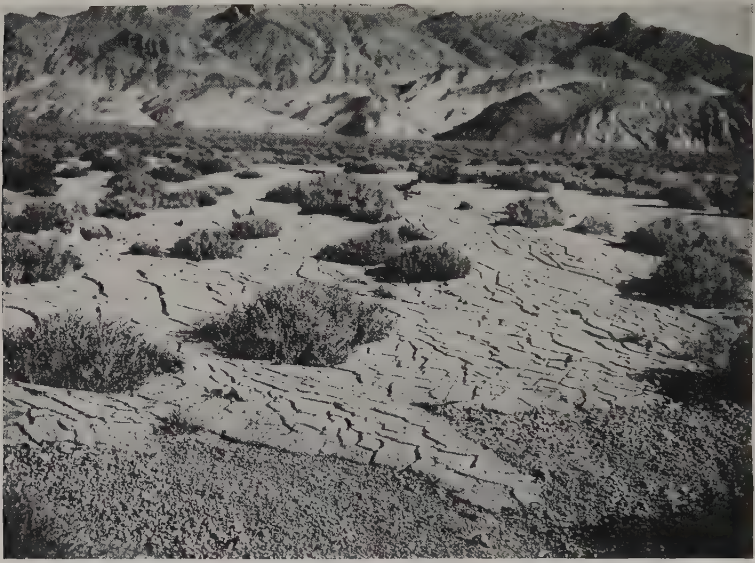


Fig. 63. — Margin of a fresh thin mudflow. East side of the Stillwater Range, Nevada. (Blackwelder.)

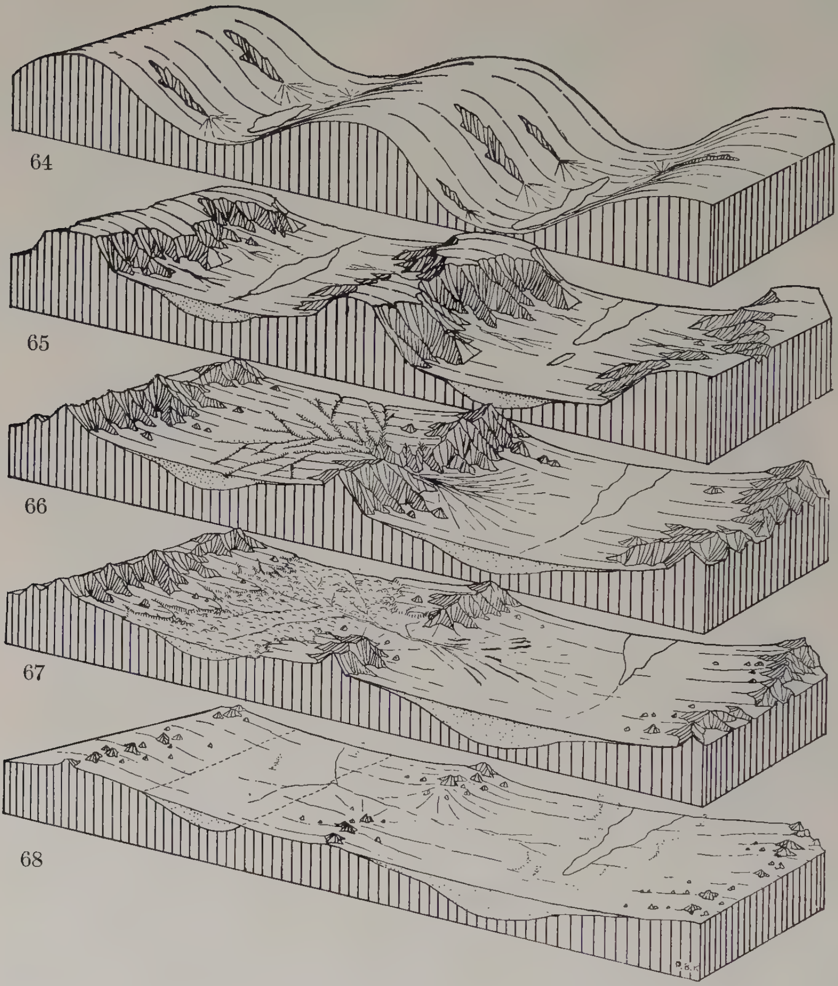
(*B*) **Mudflows.** — Another normal though infrequently operative agent of gradation in arid regions is the *mudflow.* It occurs only where earthy material becomes watersoaked on steep slopes after heavy rains, and moves downward and outward as a slippery mass (Fig. 63). It advances in waves, stopping when it becomes too viscous to flow and damming the water behind it until it liquefies again and proceeds, like an advancing flow of lava. Mudflows can carry boulders many feet in diameter. Observers have seen these great rocks bobbing "like corks in a surf." In the course of time, successive mudflows play a large part in erosion and deposition.

(*C*) **Deflation.** — The rapid mechanical weathering and scanty vegetation in arid regions are factors favorable to deflation (page 33); and in some deserts, as in the Sahara where sandstorms are frequent and violent, deflation plays an important rôle in degradation. The numerous sand dunes represent the temporary resting place of the material in transit, and ships in the South Atlantic testify to the amount of fine material blown into the sea by the prevailing winds. Deflation is not, however, of great importance in the American deserts, its effects being largely masked by the agents cited above.

Outline of the Cycle

(1) **Initial Stage.** — The following brief outline, following the studies of Davis, is organized with reference to Figs. 64–68. The conditions considered are to a certain extent special, but general principles are illustrated. Let a land surface in an arid climate be warped up into sharp folds. With the beginning of uplift, consequent streams develop, carrying drainage from the new highlands down into the adjacent troughs (Fig. 64). They flow only after violent storms and are represented at other times only by dry valleys. Most of them evaporate or sink beneath the surface before they reach the bottom of the trough, depositing their loads in rows of alluvial fans which flank the highlands. Growth and coalescence of the fans result in the narrowing of the intermont basins. The floors of these basins are called *playas* (Fig. 110), and the waters that reach them are impounded as shallow *playa lakes*, only to evaporate soon afterward. The broader term *bolson* (Spanish "purse," i.e., a pocket or basin without outlet) is commonly used to describe the whole waste-filled valley between two desert ranges. Figure 64 shows the initial surface, initial consequent valleys, and two playas in the downwarped troughs.

(2) **Youthful Stage.** — Instead of being increased as in the normal cycle, the relief is slowly diminished by the removal of waste from the highlands and its deposition on the lower slopes and in the playas. In this way two bolsons are developed (Fig. 65), their centers occupied by playas and their sides flanked by ragged mountain escarpments strongly dissected by steep gulches and valleys. The baselevel, instead of being single and fixed as in a humid climate, is multiple, being formed by the various bolson surfaces; and is moreover gradually rising



Figs. 64–68. — Ideal cycle of stream erosion under an arid climate. (Compare Figs. 46–51.)

Fig. 64. — Initial stage, showing development of alluvial fans and playas.

Fig. 65. — Youthful stage, showing decrease of relief as the bolsons rise by filling.

Fig. 66. — Mature stage, showing capture of the higher bolson by the lower one.

Fig. 67. — Later mature stage, showing dissection of the higher bolson, transfer of the waste to the lower, and the exposure of rock benches.

Fig. 68. — Old age stage, showing disintegration of the drainage, low relief, and climax of wind action.

as the deposits rise at the expense of the wasting highlands. With decreasing gradients, the mountain streams deposit farther and farther up their valleys. Thus the heads of the fans migrate slowly backward toward the mountain crests. All the streams are intermittent, flowing only after rains. A certain amount of deflation takes place, especially of the fine material on the playa flat.

(3) **Mature Stage.** — As the mountain divide between two bolsons is cut down and the bolson floors are concomitantly built up, the higher basin in time comes to drain downward *across* the old divide into the lower one. When this occurs, the mature stage is said to be reached, and drainage may pass from the upper basin to the lower in times of rain (Fig. 66). The upper basin thereby begins to be dissected by a consequent system of gullies working headward from the new channel, and the débris from their excavation is deposited as a great fan in the lower basin, hastening its filling. The surface of the lower basin has now become the master baselevel controlling both its own streams and those of its neighbor.

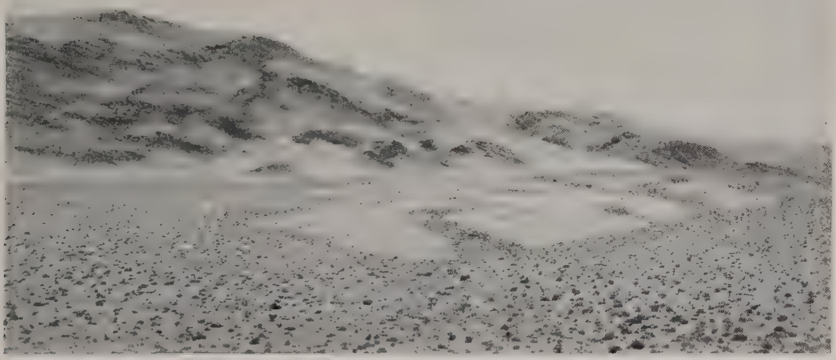


Fig. 69. — A barren desert range and the adjacent fan slope. Note the sheets of gravel spread out by flood water that rushed down the mountain slopes and quickly disappeared by sinking into the fan waste. Spotted Range, Nevada. (Erdmann.)

Full maturity is reached when the upper basin is so thoroughly dissected (usually into badlands, since they are composed of loose unconsolidated deposits) that every part of it drains down into the master basin (Fig. 67). The initial highland surface is gone, and the mountains are cloaked ever more completely with waste.

(4) **Old Age Stage.** — With the lowering of the mountains the rains, infrequent at the outset, become even more rare since condensation

decreases with decreasing relief. The whole process outlined above is correspondingly retarded, but the wind, less hampered by stream action, tends to erode hollows in the fine loose material of the basins. This breaks up the drainage pattern by forming hollows in various parts of the master bolson. The higher basin, stripped of most of its earlier mantle of waste, is largely floored with bare rock planed to a platform or *pediment* by the debouching mountain streams (Figs. 67 and 68). The wind becomes an increasingly important agent of erosion, disintegrating the drainage still further, and slowly lowering by deflation the whole surface, which now resembles a plain more nearly than a pair of basins. Only the rock masses that are most resistant to weathering remain as monadnocks. Since the wind can work at will over the entire area, the surface is slowly worn down to essentially the same level (Fig. 68). This final floor is thinly veneered with waste and dotted with monadnocks.

CHAPTER V

GLACIERS AND GLACIATION

Glaciers carry off accumulated snowfall from the lands, as streams remove the surface waters. The study of existing glaciers discloses the pronounced changes they effect upon the land, and the observation of similar effects upon lands where today there are no glaciers leads inevitably to the conclusion that glaciers must once have existed there. First, then, we shall study existing glaciers; next we shall examine the geologic work they perform, and then reconstruct some of the glaciers of the past.

The Growth of Glaciers

Perpetual Snow Fields. — On all the continents, with the exception of Australia, there are places where some of the winter snow remains unmelted from year to year, giving rise to perpetual snow fields. Such places are more numerous in high latitudes than in low, but perpetual snow fields are a familiar sight on lofty mountains, even on those that lie beneath the equator. The level above which snow is perpetual, is known as the *snow line* (Fig. 70). At the equator it lies from 15,000

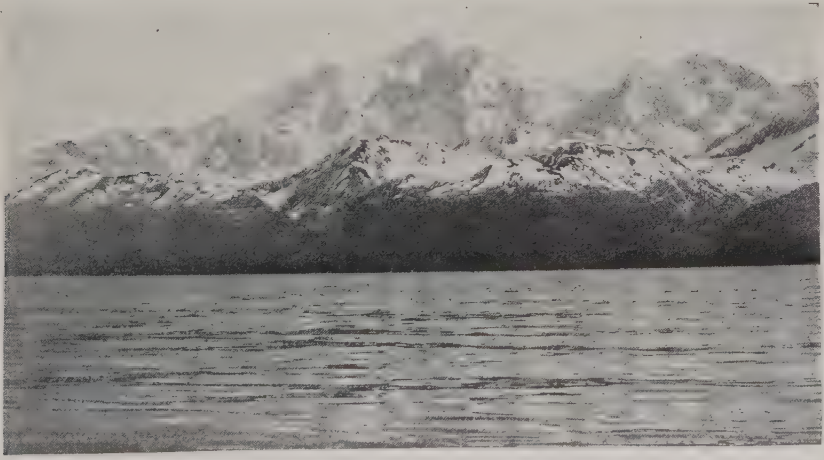


Fig. 70. — Snow line on Mt. Fairweather (15,330 feet), Alaska; from the Pacific Ocean. (Alaska Glacier Studies, Nat. Geog. Soc.)

to 18,000 feet above sea level, in Yellowstone Park 10,000 to 11,000 feet, in southern Canada about 9000 feet, in southern Alaska about 5000 feet, and in arctic America a few hundred feet. The height of the snow line is determined not only by altitude and latitude (temperature), but also by the annual precipitation, humidity, and location on the sunny or shady side of a mountain. For example, in Bolivia the snow line is 18,500 feet in elevation on the dry western side of the Andes, and 16,000 feet on the moister eastern side; and in parts of Alaska and Siberia where the ground remains frozen throughout the year the mean annual temperature is low enough for perpetual snow fields, but there is not sufficient snowfall to exceed wastage by evaporation.

Perpetual snow fields exist on most of the lofty mountain chains of North and South America. They are widespread in the higher mountains of Europe, Asia, and New Zealand; smaller ones occur even in tropical Africa. The greatest fields of snow and ice are in Antarctica and Greenland.

In southern Alaska, where the winds from the warm Japan current, heavily charged with moisture, rise over the high mountains, there is an unusually heavy snowfall, as much as 50 or 60 feet in a year. This is the greatest region of mountain glaciers in the world, both as to total area and the number of individual glaciers.

Névé; Change into Ice. — The greatest accumulations of snow occur in those parts of the perpetual snow fields whose angle of slope is less than that at which snow will slide. They are called *gathering grounds*, or, in the case of mountain snow fields, *catchment basins* (Fig. 71). They receive not only that snow which falls directly on them but also that which slides off the steeper slopes above. Under its own weight the snow becomes compacted and at the same time changes in character. The loose, feathery, newly fallen snow soon assumes a granular texture like coarse sand, and resembles the hailstone type of snow such as we see in the spring, in the remnants of winter snowdrifts. It is called *névé.* The change takes place largely as the result of alternate thawing and freezing at the surface, in the course of which the larger grains of snow grow at the expense of the smaller ones. Whole snow fields so transformed become *névé fields* or slopes.

Beneath the surface, if the thickness of snow is great, the névé becomes compacted as the air is excluded, and passes into solid ice. This ice is more or less distinctly stratified, or banded, due to successive snowfalls of somewhat different consistencies, or to the presence of wind-blown dust.

Movement; the Glacier Formed. — Before any considerable thickness of snow and ice can accumulate in a mountain region, the growing masses spread slowly outward and downward. Each mountain glacier is a live river of ice by means of which the excess snow and ice are drained off from the higher places to the lower.

The exact starting point of a mountain glacier is somewhat vague. Its upper limit is usually considered to be the snow line, but there is movement in the névé fields far above this level. In most glaciers there is a zone of prominent cracks, called the *bergschrund,* between the snow slopes and the glacier.

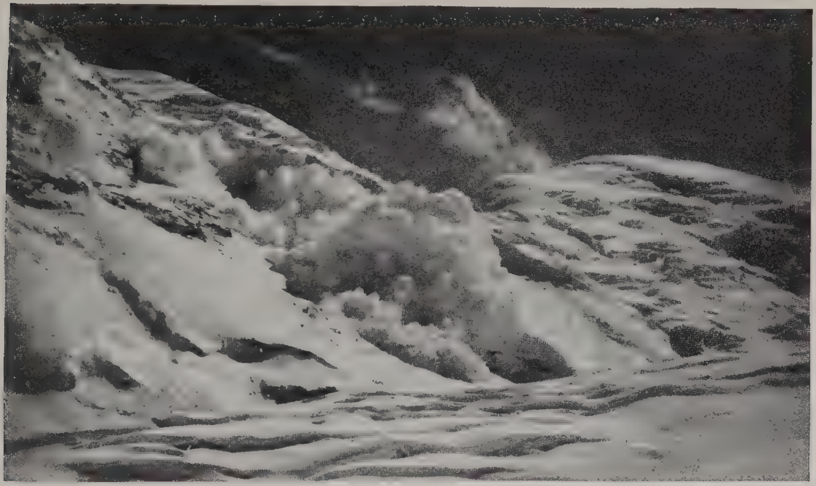


Fig. 71. — The gathering grounds of the snows. Snow and névé fields on Mt. McKinley (20,300 feet). Note the transfer of snow by a snowslide from the upper steeper slopes to the catchment basin below. (La Voy.)

Not every snow field gives rise to a glacier; some are too small to form more than a tract of névé that passes into ice beneath. Some large névé fields form at their lower ends ice masses that give evidence of some movement but do not project as ice tongues for any appreciable distance below the snow line. Such masses are called *glacierets, hanging glaciers,* or *cliff glaciers.* Most of the so-called glaciers of the Rocky Mountains and the Sierra Nevada Mountains of the United States are of this class (Fig. 72).

Lower Limit of Glaciers. — Glaciers disappear either by melting or by emptying into the sea. The distance to which a mountain glacier descends below the snow line before being halted by melting, depends upon the balance between the forward movement and melting. As an illustration the distance a rod of ice can be thrust into a furnace before being melted depends upon the size of the rod, the rapidity with which it is pushed forward, and the heat of the furnace. Thus, in warm regions, glaciers in general project but a short distance below the snow line; if a glacier is large, or flows with relative rapidity, it will extend farther than under the opposite conditions. In moist regions, where there is abundant snowfall, the glaciers are larger, flow more rapidly, and therefore descend farther, than in dry regions at the same latitude. As we go to higher latitudes, the glaciers reach lower elevations; in sub-arctic regions they actually reach the sea and break off to form icebergs.

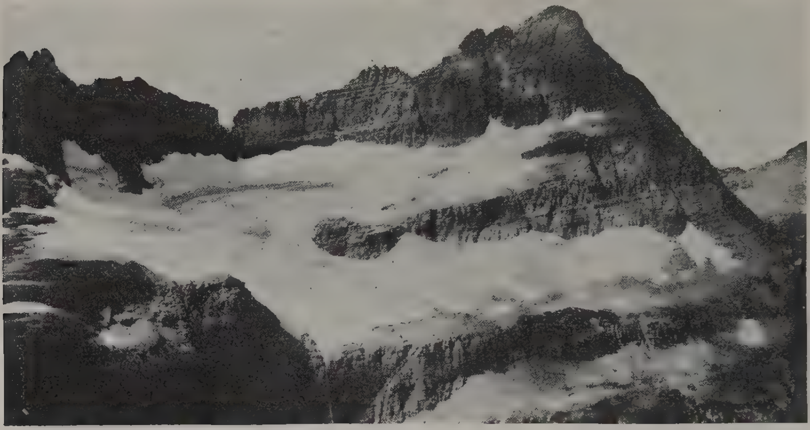


Fig. 72. — A glacieret. Shepard Glacier, Glacier National Park. (Alden.)

In the Alps and in Norway, glaciers project as far as 5000 feet below the snow line. In southern Alaska they reach sea level at about 55° N. latitude, and in southern Greenland at about 60°; whereas in Norway at 70° N. they melt before reaching the sea. In southern New Zealand at latitude 45° S., glaciers descend into luxuriant forests, and in Chile, glaciers from the Andes reach the sea at about 47° S.

Classes or Types of Glaciers

Glaciers are classified according to their shape, size, and location, as: *valley* or *alpine* glaciers; *piedmont* glaciers; and *ice caps.* Enormous ice caps, called *continental* glaciers, overran the northern countries in the recent geologic past, but are known only indirectly through their effects on the land surface. An understanding of them will be made clearer by considering first the separate types of existing glaciers.

Valley Glaciers. — It is valley glaciers that are usually thought of when glaciers are mentioned, for they are the common kind the world over. They are fed by mountain snow fields and flow down existing valleys. Like rivers, they follow the valley windings and spread from side to side, and commonly they have tributaries (Fig. 73) that join the master ice stream and swell its mass. Furthermore a glacier may split about islands and come together again downstream as an unbroken mass of ice. But the resemblance to rivers is largely superficial; differences are more pronounced than are similarities.

There are about 2000 valley glaciers in the Alps. It was here that they were first studied, and that is why they are often called alpine glaciers. Most of them are less than 2 miles in length; a few are from 3 to 5 miles long, and one, the Great Aletsch, is nearly 10 miles long. Similar valley glaciers may be seen in other parts of Europe, — in Norway, the Pyrenees, and the Carpathians. Magnificent glaciers, up to 30 miles in length, are to be found in the Himalayas, and other high ranges of Asia furnish fine examples. Even under the equatorial sun of Africa there is a small glacier on the slopes of Kilimanjaro (20,000 feet). In the United States hanging or cliff glaciers are numerous, but true glaciers lie only on some of the lofty volcanic peaks in the west, such as Mounts Rainier, Shasta, Hood, and Baker. Those on Mount Rainier, in Washington, attain a length of 7 miles. The coastal mountains of British Columbia contain many fine glaciers. In southern Alaska valley glaciers exist in unknown thousands, or perhaps tens of thousands, nourished by great unexplored snow fields that mantle the lofty mountains. Some attain 50 miles in length and 5 to 6 miles in width, and descend from heights as great as 18,000 feet above sea level. Scores of them enter the sea and give rise to icebergs.

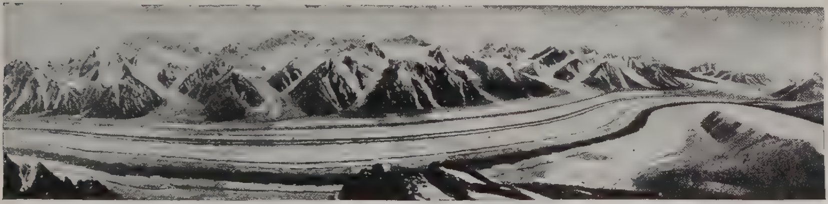


Fig. 73. — Typical valley glacier with branches; the Siachen Glacier, Himalaya Mountains. Moraines of earth are seen on its surface as dark bands. The part of the main glacier shown in the view is nearly 30 miles long. Note the steep tributary glaciers descending from the high snow fields.

Piedmont Glaciers. — The name implies a glacier at the foot of a mountain. It designates those glaciers that have descended from the mountains and have spread bulb-like upon the gentler sloping plains beneath. The former expanded foot of the Rhone Glacier (Fig. 77) was an example on a small scale. Usually several valley glaciers coalesce to form a piedmont glacier. If a valley glacier be likened to a river, then a piedmont glacier may be compared in size to a lake. Such glaciers are not common, and are confined to high latitudes; the Malaspina and Bering glaciers of Alaska are examples.

The great Malaspina Glacier lies at the base of the range that supports Mt. St. Elias (18,000 feet) and Mt. Logan (19,540 feet), and spreads over an area of 1500 square miles to the edge of the sea (Fig. 74). It has a gently rolling surface, broken by innumerable fissures,

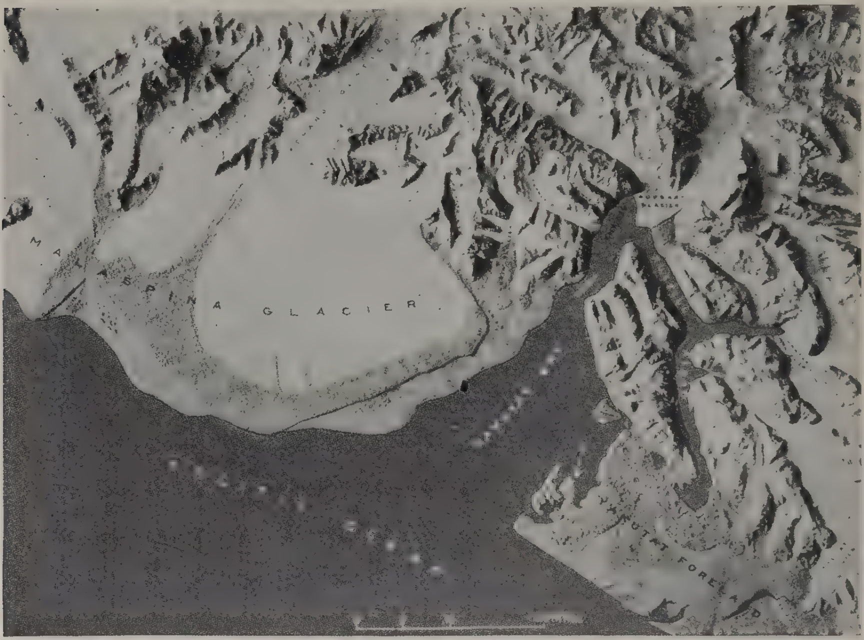


Fig. 74. — Model of Malaspina Glacier, Alaska. (Martin, Univ. of Wisconsin.)

irregular hummocks, and piles of débris. Its borders are so mantled by dirt that they support a dense forest which rests on 1000 to 1500 feet of ice. The melting ice nourishes several short rivers of large volume.

Ice Caps. — These are vast sheets of ice which, to carry the analogy further, resemble seas in size. The picture they present is that of an almost endless monotony of desolate, wind-swept ice, gently rolling, and with occasional mountains of rock, known as *nunataks*, projecting through like islands. At their borders ice caps taper down to pronounced lobes, or to tongues of ice that discharge into the ocean, forming icebergs.

Only two large ice caps exist today, in Greenland and Antarctica; smaller ones occur in Iceland. The Greenland ice cap is about 1300 miles long and has an area of about 715,000 square miles. It has now been crossed several times, notably by the explorer Koch. He states that there are two great flattish domes from which ice spreads out in all directions. The northern dome reaches an altitude of nearly 10,000 feet and the southern about 8500 feet. The thickness of the ice probably ranges from 2000 to 7000 feet. The edge of the ice is definitely known to be in motion locally, since it discharges into the sea. In the northern part of Greenland, however, where the snowfall is light, the ice is thin and stagnant.

The Antarctic ice cap is thought to have an area of 5,000,000 square miles. Its interior has been partly explored by Shackleton, Amundsen, Scott, and Byrd. Its thickness is unknown but its surface reaches an altitude of about 12,000 feet. According to Scott, it pushes off the land out over the sea to form vast stretches of floating fields of ice, known as the " Great Ice Barrier." The huge tabular icebergs of the Antarctic break off from the barrier ice.

The Movement of Glaciers

That glaciers actually move had long been suspected but was not generally known until 1827, when Hugi built a hut on the Aar Glacier, and its change of position was observed. Since then many accurate measurements have shown not only the rate but also the nature of the movement. The movement is always downstream, even though a glacier is said to be retreating; a " retreat " of a glacier is of course not a bodily movement of the ice back towards its source, but simply a retreat in the position of the ice front owing to excess of melting over forward motion.

Rate of Movement. — Hugi's hut moved down the Aar Glacier a distance of 4650 feet in 15 years; 44 years later it was found 7900 feet farther down the valley, indicating a rate of movement of from 6 to 10 inches per day. A similar long-time measurement came to light for the Glacier des Bossons on Mont Blanc. In 1820 three guides were buried beneath an avalanche on the mountain. It was predicted by Dr. Forbes in 1858 that their bodies would be given up by the glacier 35 to 40 years after their burial; and just 41 years later the heads of the three guides, with some hands and clothing, appeared at the foot of the glacier. The average rate of movement was 8 inches a day. Other measurements have been made by placing markers on the ice and noting the time they have taken to travel a measured distance.

Such measurements in the Alps have shown that the rate of movement there seldom exceeds 1 to 2 feet per day. The large Muir Glacier in Alaska was found to have a motion, at its middle, of 7 feet per day. Rates as high as 60 to 75 feet per day have been recorded in some of the tongues of ice that move toward the sea through narrow mountain passes from the Greenland ice cap.

The rate of movement is influenced by several factors: it is greatest in those glaciers that have a large supply of ice; it increases with a steep valley slope and a smooth bed; it is greater in summer, when there is more water in the ice, than in winter; it increases when the temperature of the ice approaches the melting point. The gradient, the amount and thickness of ice, and the temperature, are thus the chief factors that affect the rate of movement.

Differential Movement. — An important discovery in regard to glacier motion is that the rate of movement is not the same in all parts of its mass. A glacier does not move as a whole by sliding down its bed, like a cake of ice off the roof of a house. It has been found that stakes driven in a straight row across the top of a glacier after a time become curved downstream, proving that the middle portion moves faster than the sides. Similarly, vertical lines of pegs on the side of a glacier show that the top moves faster than the bottom. Also, it has been observed that the line of swiftest motion does not everywhere lie in the middle of the glacier, but is sinuous. The movement of the ice is therefore *differential;* that is, some parts of it move faster than other parts. This conclusion is important to a proper understanding of the cause of glacier motion.

Nature of Glacier Movement. — The observation that glacier movement is differential led early observers to assume that the flow of glacier ice is similar to that of a stiff, viscous fluid, such as pitch or asphalt. This, however, cannot be the case; the viscosity is only apparent, and not real. In a liquid, the molecules are free to move readily in any direction, but in a crystalline solid the molecules or atoms are fixed in definite positions from which they can be moved only by great force. Ice is a true solid; its component particles are definitely arranged with respect to each other in a geometrical pattern. Since it is crystalline, it cannot flow by free movement of the molecules and therefore cannot exhibit the property of a liquid. Furthermore, a glacier will not turn and flow up an empty tributary valley, even though the top of the ice may tower high above the tributary floor; the Kennecott Glacier in Alaska forms a dam several hundred feet high across Hidden Creek tributary, but the ice does not flow into Hidden Creek valley. Ice must flow as a solid; it does so because it becomes plastic under stress, and so resembles a viscous substance. It is a weak solid and yields under the enormous weight of the glacier. An analogy is found in the plastic behavior of certain metals under stress. One need only deform a piece of lead or copper to be aware how small a force is required to make these weak metals flow.

VARIOUS FEATURES OF EXISTING GLACIERS

The surface of a glacier is not an expanse of smooth ice; typically it is highly irregular, so that travel over it may be laborious and hazardous. It is traversed by wide and deep cracks called *crevasses*, and is covered in places by accumulated heaps of earth and rock fragments called *moraines.* Unequal melting also gives rise to a variety of relief forms that hinder the traveler.

Crevasses. — Among the most striking features of a glacier are the great yawning cracks or crevasses in its upper surface. Many of these are 20 feet or more in width and up to 300 feet in depth, and they occur anywhere between the névé fields and the lower end. They are formed by tension acting on the brittle ice, and are induced where the ice rides over an uneven bed, or where one part moves faster than another part. Their abundance varies with the amount and the rate of the straining. *Transverse* crevasses are most prominent where a glacier flows over a sharp incline in its bed, giving rise to an *ice fall.* Where a glacier emerges from a constricted to a wider part of a valley, *longitudinal* crevasses occur, because the glacier tends to spread sideways and the ice is pulled apart. These are particularly characteristic of the terminal lobe of a glacier, as in Fig. 77. *Marginal* crevasses commonly occur along the sides of a glacier. These usually point inward and upstream at an angle of about 45° to the course of the glacier. They are formed as a result of tension set up by the central part moving faster than the marginal parts.

Crevasses, once formed, do not remain open indefinitely. Not uncommonly, however, the upper parts of crevasses are widened by melting so that slices of ice between parallel cracks become sharpened to thin blades and needles, called *seracs.* These persist long after the lower parts of the crevasses are closed, and a considerable part of the glacier may present a maze of sharp ridges such as one meets in trying to cross the Mer de Glace. On some large valley glaciers, such as those of Alaska, these ridges take the form of steep hillocks 100 to 200 feet high.

Moraines. — In a river valley, the débris that is worn from its sides forms local talus slopes or is washed down into the stream and borne away. But in a valley occupied by a glacier, this material descends upon the ice to form a marginal fringe of débris that is carried along slowly by the moving ice. Thus there are formed continuous bands of débris along the margins of the glacier; these are called *lateral moraines* (Fig. 75). Where tributary glaciers enter, *medial moraines* are formed (Fig. 75). Ordinarily a glacier has several medial ribbons of débris, each separated from the others by "white ice," and the number present usually indicates the number of tributaries that have joined the master glacier (Fig. 73). As the glacier reaches lower levels and undergoes surface melting, the ice beneath the débris melts less rapidly, so that commonly the moraines are prominent ridges up to 100 feet in height. The débris will slide down the sides of such ridges and so widen the morainal bands.

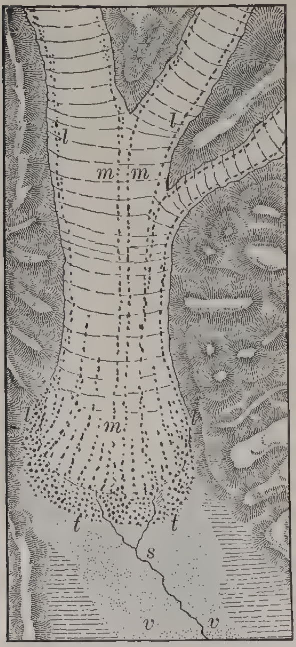


Fig. 75. — Plan of a valley glacier, showing tributaries and terminal lobe; *ll*, lateral moraines running into *tt*, the terminal moraine; *mm*, medial moraines. As melting progresses, more and more material appears; *s*, exit of subglacial stream; *vv*, valley train of water-laid débris.

The débris is not only on top of the ice (*superglacial*) but is also within the body of the glacier (*englacial*). Much of the latter is carried along near the bottom of the glacier. The englacial material is derived partly from débris that fell on the surface as the glacier was being built up; partly from material that fell into the crevasses; and partly from material that was scraped off the sides and bottom of the valley. Much of the englacial material becomes superglacial where rapid melting occurs near the end of the glacier, and, in combination with spread-out lateral and medial moraines, may cover the entire "snout" of the glacier. The transported material that is eventually dumped at the end of the glacier, in a confused mass of earth and stones, forms the *terminal moraine.*

Glaciers are thus great transporters of débris; fine dust, and boulders of thousands of tons weight are carried with equal facility.

Differential Melting. — The presence of débris on the surface of a glacier causes uneven melting by the sun, and numerous irregularities result. Thin layers of dark débris or dust on the glacier absorb more heat from the sun than does the adjacent white ice, and on becoming warmed through melt the ice beneath to form *dust wells* or " bath tubs." If the débris layers are thick, they act as heat insulators; the ice beneath is protected from melting and prominent ridges or *dirt cones* are formed. Some large slabs of rock eventually cap ice pedestals (Fig. 76) and are known as *glacier tables.*

Drainage. — If one walks over a glacier on a warm day, he hears the murmur of innumerable rivulets of water that are formed from the active melting that goes on. Many join together to make streams so wide that they are difficult to jump across, and these plunge down deep crevasses, commonly into circular shafts called *moulins* or *mills,* of their own making. And deep beneath is the subdued roar of larger subglacial streams that flow in ice tunnels. All along the sides, also, numerous trickles of ice water descend and commonly form marginal streams. In a warm summer many feet are melted from the top of a glacier. On one small glacier in Alaska, telephone poles sunk 6 feet in the ice had to be reset twice during the summer.

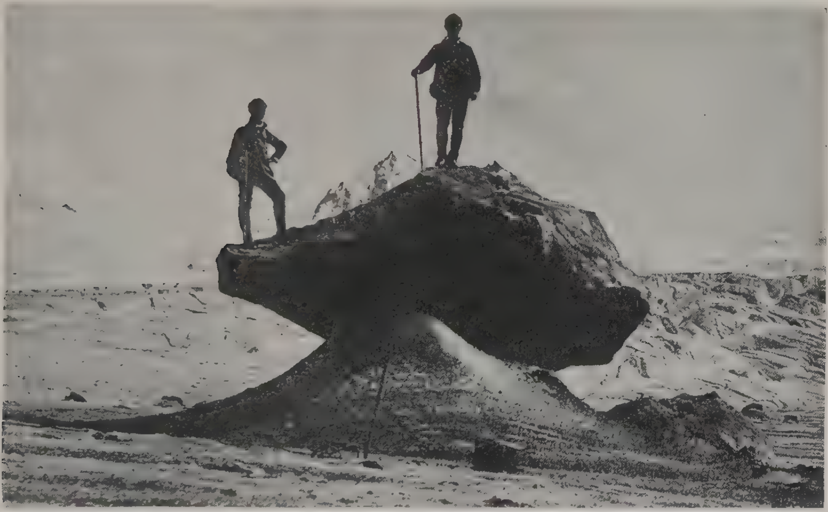


Fig. 76. — Glacier table mounted on ice pedestal; caused by differential melting. Switzerland.

The subglacial streams emerge from caves at the front of the glacier or less commonly they spurt forth from large vertical holes, and form the starting points of good sized rivers. The water is striking in appearance, for it has a peculiar opaque grayish-white color that has given rise to the name *glacier milk.* This is due to the presence of fine unweathered rock material that has been ground from the glacier bed and is carried in suspension. This milky water can be traced far down the valley or out to sea, and its presence makes the traveler suspect that he is approaching a region of melting glaciers.

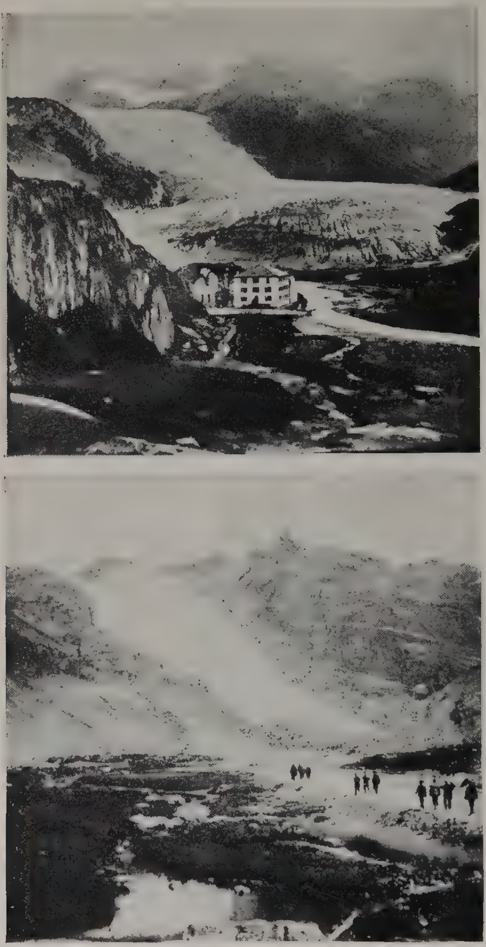


Fig. 77. — Two views of the Rhone Glacier, Switzerland; the upper one taken in 1870, the lower in 1905, illustrate the retreat of the ice in 35 years. The older one shows the terminal lobe and longitudinal crevasses.

Advance and Recession of Living Glaciers. — Glaciers appear to advance and retreat in response to varying climatic cycles of greater or less precipitation, or of sunshine and cloudiness. At present there seems to be a general condition of recession, although there are at the same time minor oscillations of retreat and advance. In Europe and North America glaciers have been retreating, in general, for several decades. The Muir Glacier in Alaska has retreated more than 12 miles in the last 30 years. The amount of retreat of the Rhone Glacier is shown in Fig. 77. In the Chamonix Valley in the Alps, however, the glacial retreat from 1812 to 1910 amounted at most to only ½ mile.

On the other hand, some individual glaciers have shown prominent advances. For example, all the glaciers on the Savoy side of the Mont Blanc chain advanced prominently between 1910 and 1920. The Grindelwald Glacier, following a long retreat between 1875 and 1900, had by 1924 advanced a considerable distance. In Alaska, a lobe of the Childs Glacier advanced 2000 feet in 1910 to 1911, receded for the next few years, and advanced a short distance again in 1925. These advances appear to be only temporary oscillations, although in the case of the Alaskan glaciers the length of time over which observations have extended is too short to state definitely whether there is a general retreat or advance.

EROSION BY MOUNTAIN GLACIERS

Valley glaciers by their erosion give rise to distinctive floor features and characteristic valleys, and at the same time those parts of mountains that project above the glaciers are profoundly modified. How these changes are brought about will become clear with an understanding of the nature of glacial erosion.

Nature of Glacial Erosion. — Glaciers erode by *plucking* and by *abrasion*. In plucking, the glacier actually quarries out masses of rock, incorporates them into itself, and carries them along. This is accomplished by water that trickles down into crevices in the rock, freezes, and springs out blocks of rock; or that freezes around blocks bounded by joints or crevices so that they become a part of the body of the glacier; or by the ice of the glacier being pressed into the cracks in the rock by its own weight. As the glacier moves forward, the blocks are plucked out and dragged along with it. It is a powerful quarrying

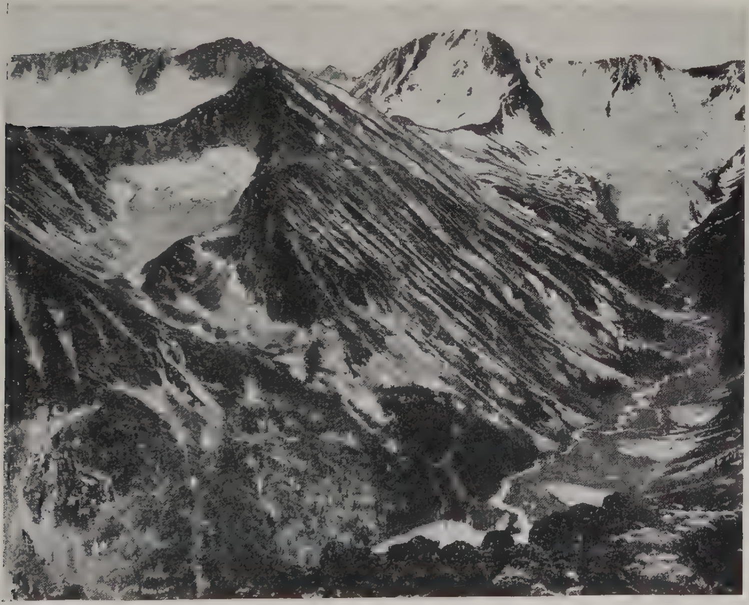


Fig. 78. — A group of glacial cirques (amphitheaters); also a small valley glacier, and a glacial stream and lake. Coast Range, British Columbia. (Bateman.)

operation. At the heads of mountain valleys this process gives rise to bowl-shaped basins called *amphitheaters* or *cirques* (Fig. 78). Commonly these are cut somewhat deeper at the center than at the place of discharge, and when the glaciers disappear, these depressions form some of the beautiful mountain lakes or tarns (Fig. 79), such as Lake Louise in the Canadian Rockies. Such cirques are common features of high mountains in middle and high latitudes, and are evidence of the former existence of glaciers. The process of plucking operates also on the sides and bottoms of a valley glacier, particularly where projections of rock extend into the ice and where the bedrock is much jointed.

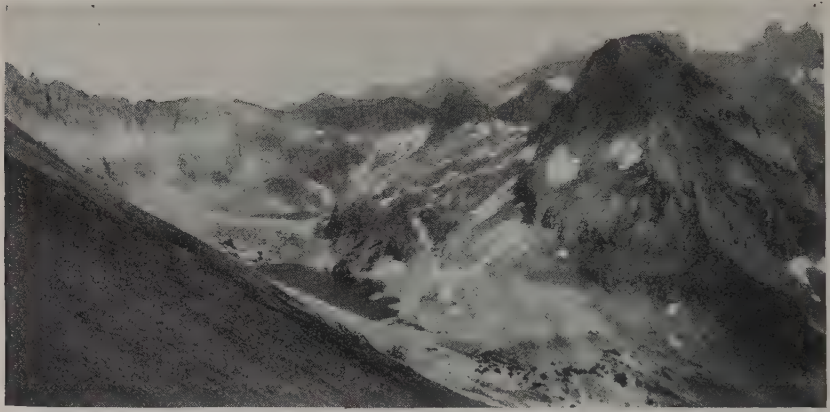


Fig. 79. — Lion Lake and another small tarn in a glacial cirque south of Triple Divide peak, Sierra Nevada, Calif. (Matthes.)

Dirt and stones frozen fast in the bottom and sides of the ice form a huge slow-moving rasp that grinds away or *abrades* the bedrock of the bottom and sides of the valley. As the cutting tools wear out, new ones take their place. This rasping power is enormously augmented by the great weight of the overlying ice.

The rate of erosion is influenced by various factors. It is obvious that the larger the glacier, the greater will be its erosive power, because of its great weight; consequently master glaciers erode their valleys more than do smaller tributary glaciers. A thin ice sheet may override soil without even removing it. Also, clear ice produces no abrasion; the greater the amount of débris in the ice, up to certain limits, the more it can grind the bedrock. Erosion is favored by fast movement of the glacier and by weak bedrock. Another factor is the firmness with which the teeth of the rasp are gripped by the ice; where the ice is soft and melting, as at the lower ends, it may not even remove loose soil. Plucking plays a small part if the bedrock is smooth and unjointed, but becomes highly effective if the rock surface is rough and jointed.

In contrast to a river, a glacier can cut far below grade locally and scoop deep hollows in bedrock along its course. As a result, glaciers erode valley bottoms unevenly; pronounced deepenings of a glaciated valley occur in constricted parts, where the ice has been forced to move faster, or in places underlain by exceptionally weak rock.

Results of Erosion. — Results of glacial erosion are evident in bedrock features, in the character of the valleys, and in superglacial effects.

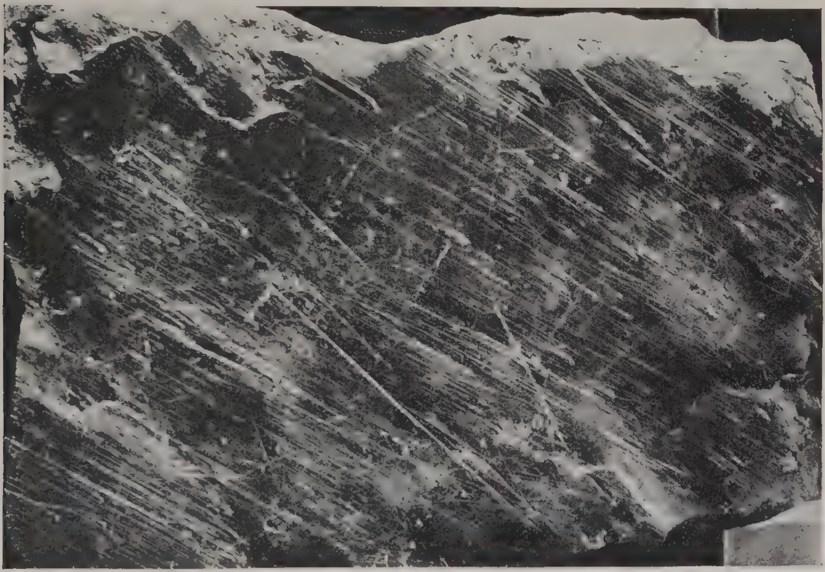


Fig. 80. — Glacial striae; scratches and groovings made by moving ice on limestone bedrock. Near Rochester, N. Y.

The Valley Floor. — If one walks up a valley recently vacated by a glacier, he will notice that rock outcrops on the bottom and sides are smoothed, rounded, and scored by scratches and grooves, called *glacial striae*, that trend in the direction of glacial flow (Fig. 80). These marks were caused by the rock-shod bottom ice that dragged heavily over the floor; the pebbles and boulders scratched and grooved, and the dust polished. The rock flour thus formed was swept away by the subglacial drainage and gave rise to the "glacier milk" previously referred to. The traveler may have to circle around many ponds or pools that occupy gouged-out rock basins in the valley floor. If he glances down the valley, he sees projecting from the floor numerous smoothed, rounded rock masses, elongated in the direction of flow, which, following the French terminology, are called *roches moutonnées* (Fig. 81). The presence of all these features is a clear indication of the former existence of a glacier.

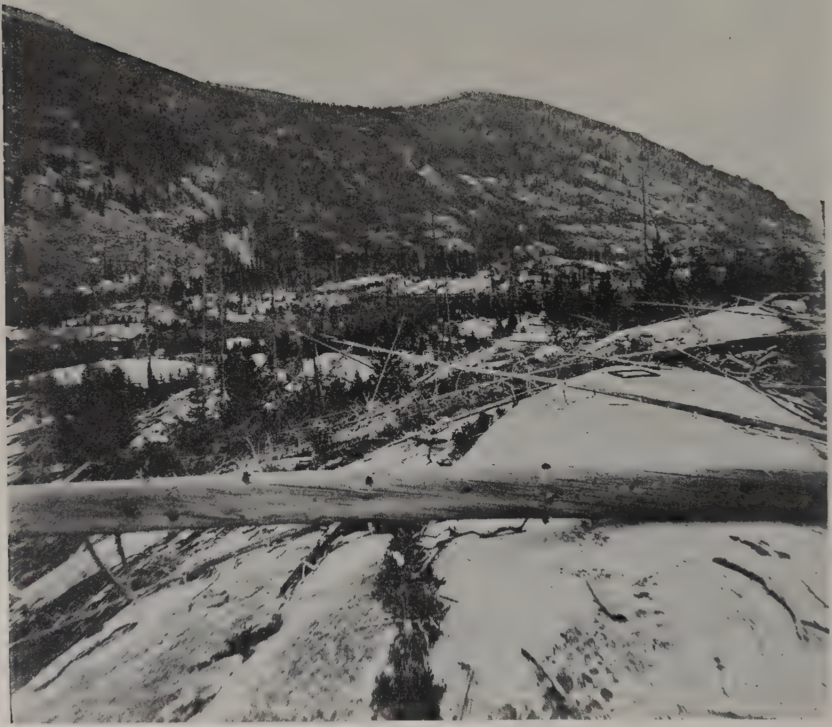


Fig. 81. — Roches moutonnées near Mount of the Holy Cross, Colo. (U. S. Geol. Surv.)

The Glacial Valley. — The shape of the valley is also a distinctive feature of erosion by mountain glaciers. It has a characteristic **U**-shape, flaring at the top, with steep sides, and a broad bottom (Fig. 82). No river erosion could produce such a valley. A comparison of Figs. 82 and 84 (*a*) shows the contrast between a glacial valley and a normal youthful river valley. This unusual shape results from the fact that a glacier erodes the sides as well as the bottom of its valley, and cuts away the ends of interlocking spurs developed between the tributaries of a stream-eroded valley.

The tributaries also are peculiar. They do not join the master stream at grade as they normally do in river valleys, but commonly enter hundreds, or even a thousand or more feet, above the main valley floor. The streams that flow along such *hanging valleys*, as they are called, when they reach the abrupt drop, cascade down to the main valley floor and give rise to some of the most beautiful waterfalls (Fig. 83), such as those of the Yosemite Valley in California and the Lauterbrunnen in Switzerland. The hanging valleys originate because the large master glaciers erode their valleys more rapidly than do the smaller tributary glaciers. Consequently, the master valley becomes overdeepened by bottom erosion with respect to the tributary valley, and also the master valley becomes overwidened by lateral erosion, causing the mouth of the tributary to retreat to a higher level. Commonly there exists an abrupt change in slope between the steep glacial valley wall and the unglaciated country above. In Switzerland these "shoulders," as they are called, are usually 1000 feet or so above the valley bottoms and are a favorite place for villages. In Alaska they reach heights of 3000 feet or more above the valley floors.

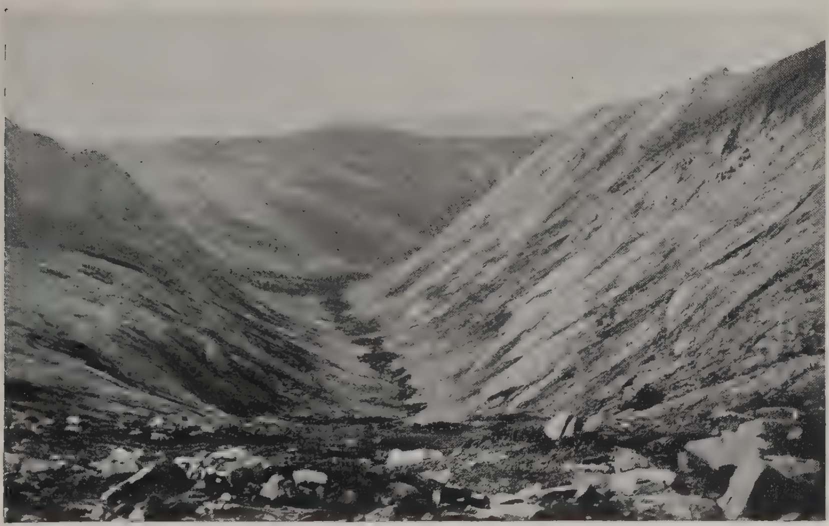


Fig. 82. — U-shaped glaciated valley. Deadman Canyon, Sierra Nevada, Calif. (U. S. Geol. Surv.)

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Other distinctive features of a glacial valley are truncated or *faceted* spurs between adjacent tributaries. They are spurs of normal river valley erosion that have had their "noses" cut off by erosion of the glacier in the parent valley. Their characteristic triangular form may be seen in Fig. 84, (*b*) and (*c*). Thus **U**-shaped sections, cirques, rock basins, hanging tributary valleys, shoulders, and faceted spurs are characteristic erosional features of glaciated valleys. The changes effected in a normal river valley by glaciation are seen in Fig. 84.

Fiords. — Where overdeepening in glacial valleys along a coastal region has extended beneath sea level, or where the coast has been depressed, the sea occupies the valleys after the ice has vanished, and *fiords* result. Most of the fiords of Alaska, British Columbia, Labrador, Chile, and Norway probably have been formed in this way.

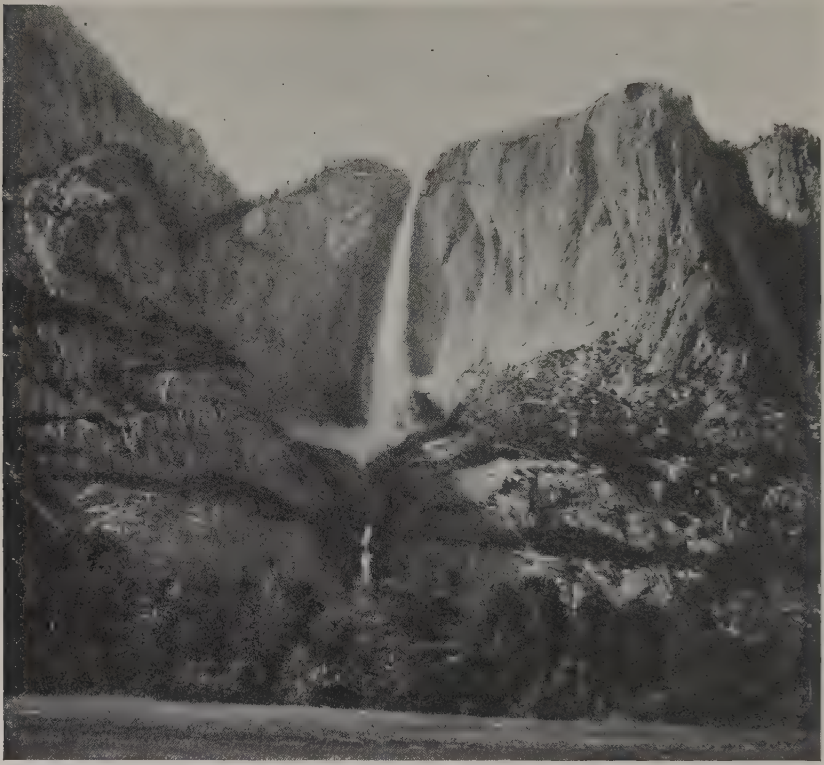
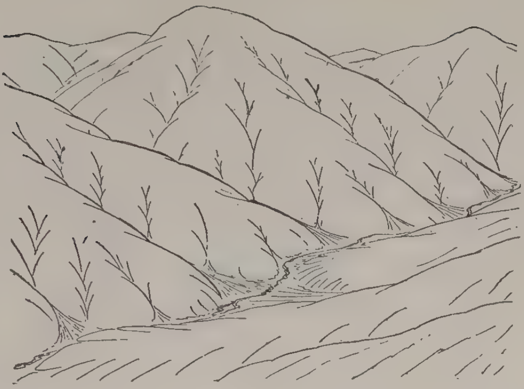
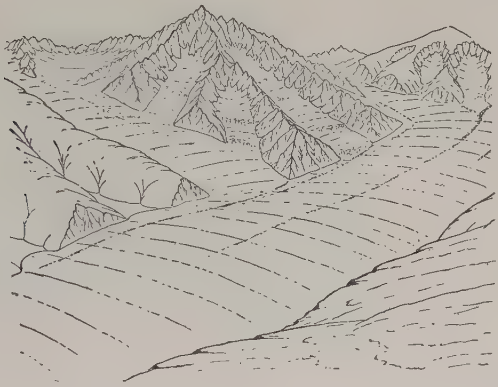


Fig. 83. — A hanging tributary valley, from which the stream falls precipitously to the main valley. Yosemite Falls, Yosemite Valley, Calif. (U. S. Geol. Surv.)

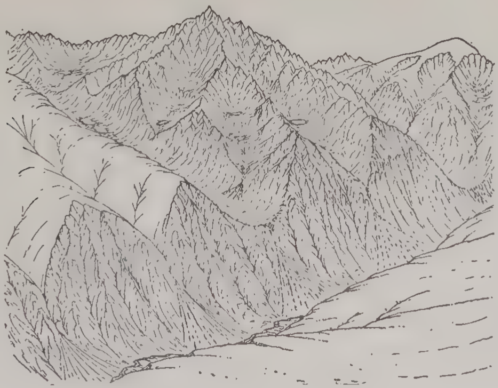
Superglacial Erosion. — The higher parts of mountains that rise above the glaciers and are covered by perpetual snow-fields are also greatly eroded while glaciation is in progress. This is shown graphically in Fig. 84, (*a*) to (*c*). Parallel tributary glaciers in the process of widening their valleys sap the sides of intervening ridges and leave jagged



(*a*) A mountain mass, normally eroded by weathering and running water and unaffected by glacial action. The valleys and ravines are V-shaped in section.



(*b*) The same mass, strongly affected by glaciers which occupy the valleys. Note the rugged topography above the ice, produced by weathering and frost.



(*c*) The same mass after the retreat and melting of the ice. Note the nature of the topography, and trough-like form of the glaciated valleys, the amphitheaters, some with lakes, the hanging valleys, and the faceted spurs. (W. M. Davis.)

Fig. 84.

comb ridges or *arêtes*. The headward gnawing in cirque formation causes cirques to approach each other from opposite sides of a ridge until they intersect, leaving jagged sawtooth pinnacles between them (Fig. 78). Or, several cirques disposed about a single mountain and eating headward, as on the right hand mountain in Fig. 84 (*b*), will coalesce until an irregular or pyramidal *horn* is the result. Such horns are typical of glaciated mountains; the famous Matterhorn in the Alps and Mount Assiniboine in the Canadian Rockies are outstanding examples. Changes above the glaciers thus result chiefly from sapping and undermining.

DEPOSITION BY MOUNTAIN GLACIERS

The vast amount of material transported by glaciers must be deposited when the ice melts. The materials transported and deposited by a glacier are called glacial deposits, or *glacial drift*. But the drift is of two distinct kinds: — one, which is dumped in heterogeneous fashion as the ice melts, and is unstratified, is called *till*, or *boulder clay;* the other is material that is sorted out and stratified by water from the melting ice, and laid down to form *stratified drift*.

Unstratified Deposits

The outstanding characteristic of glacial till is that it is unstratified and in this respect it differs from the nicely assorted water- and wind-laid sediments. Coarse boulders and fine sediments occur jumbled together just as they were dropped by the melting ice. Some of the pebbles and boulders are as unworn and angular as when they were removed from their original positions. Those that were dragged against the bedrock on the bottom or side of the valley, however, have characteristic smooth flat surfaces or *facets* (Fig. 85) and are polished and scratched. No other pebbles or boulders are quite like them; they are diagnostic of glaciation.

Moraines. — When a glacier melts, the débris contained in the ice is dropped as a till sheet or *ground moraine* over the valley bottom. Commonly it is not very thick, and much of the finer material may be washed away, leaving the boulders as the most conspicuous part of the moraine.

The lateral moraines of the glacier are left along the sides of the valley as long ridges of till, some of which are several hundred feet in height. The name *lateral moraines*, used for these features on the sides of live glaciers, is retained for the ridges after the glacier has disappeared.

In places the ridges are sharp-topped, because the material fell on each side to its angle of repose as the ice melted. The medial moraines of the glacier less commonly persist as distinct medial ridges on the ground; they are likely to be washed away, owing to their central position in the valley. In favored places, however, parts of old medial moraines are well preserved.

The *terminal moraine* consists of jumbled masses of till, usually of irregular shape, extending across the valley. Commonly it has a general crescentic form, concave toward the glacier. It is composed of pebbles and boulders, faceted and angular, heterogeneously admixed with earthy materials. The piles range in height from a few tens to a few hundreds of feet. The higher ones usually indicate that the ice front was relatively stationary for a long time so that much débris was deposited at one place. If the ice front in its retreat makes extended halts at intervals, a series of *recessional moraines* are left. The terminal moraine is usually breached and more or less destroyed by streams. The surface of a typical terminal moraine is characterized by numerous rounded mounds and small undrained basins, with a haphazard distribution. This peculiar kind of surface, known as *knob-and-basin* topography, is caused by highly irregular deposition of the terminal moraine. This irregularity results from frequent minor fluctuations of the ice front and from variation in amount of débris delivered by the ice to any one point. Some of the basins in an abandoned terminal moraine form small ponds.

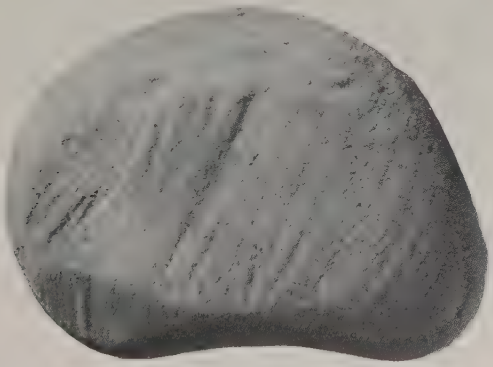


Fig. 85. — Faceted and striated pebble. Sierra Nevada, Calif. (W. D. Johnson.)

Glacial Boulders, or Erratics. — Another characteristic feature of a glaciated valley is the presence of scattered boulders, or *erratics* (Fig. 86), of all shapes and sizes, many of which are foreign to the underlying rock; boulders of granite derived from the head of the valley may rest on limestone in the lower part of the valley. Some of them are perched in insecure positions (*perched boulders*). These erratics will be referred to again under continental glaciation.

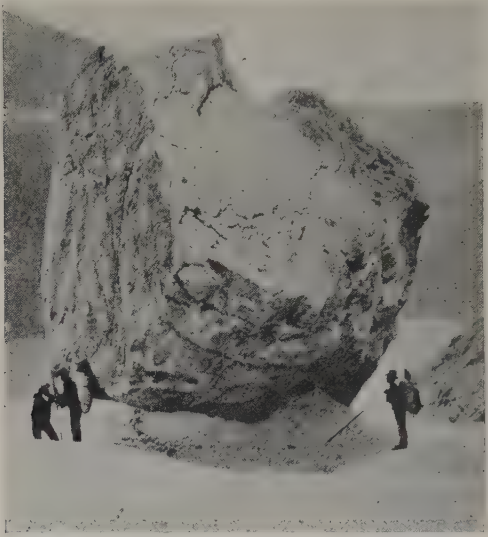


Fig. 86. — Erratic block, Alaska boundary. (Coleman, after Melson.)

Stratified Drift

The streams discharged by melting glaciers carry out boulders, pebbles, sand, and silt. The boulders, however, are quickly dropped; the remaining materials build up an *outwash plain*, or a *valley train* which may extend far down the valley. Nearest the glacier are the coarser pebbles, and farthest away are the sands or coarser silt; much of the finest rock flour of the milky water is carried far beyond the valley train. Thus the deposits are sorted out by the water and are stratified into beds of sand and pebbles. The latter are somewhat rounded by the friction they experience as they are rolled along. Across these gravelly plains, where the gradient is low, the streams pursue a braided course and are continually shifting their channels. They thus work over the material and search out the finer sands to carry them farther downstream. As the outwash material accumulates it may lap up over the end of a stagnant glacier, or bury large isolated blocks of ice. Later, when the ice melts, pits or *kettle holes* result, and the stratified deposits form a *pitted outwash plain* (Fig. 87).

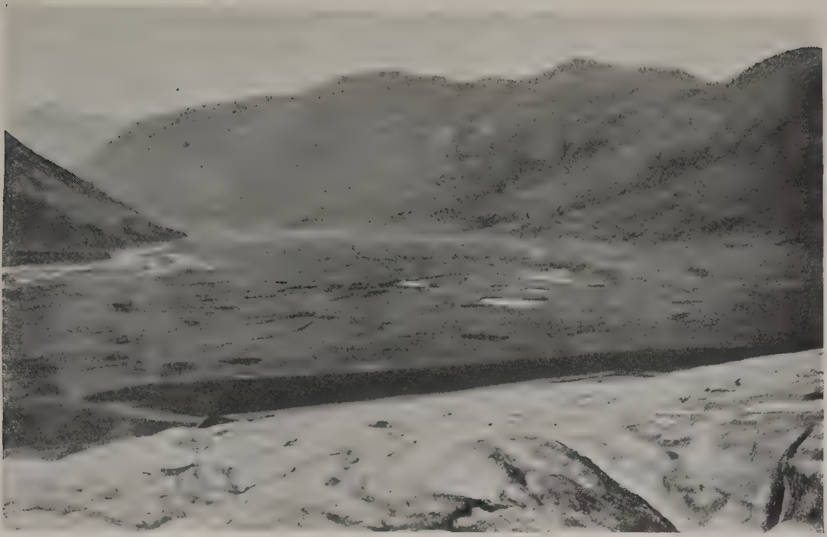


Fig. 87. — Pitted outwash plain, derived from Hidden Glacier, Alaska (view downstream). Part of ice near front of glacier shows in foreground. Note the braided streams, which are building up the plain. (U. S. Geol. Surv.)

CONTINENTAL GLACIATION

Continental glaciation on a grand scale has taken place over large areas in high-latitude lands. The glaciers themselves have largely vanished; their former presence is inferred from the results they achieved, and their character is arrived at by inference and by analogy with existing ice caps. Their action was different from that of valley glaciers; they did not flow as tongues of concentrated erosive power in valleys, and therefore glacial valleys, hanging valleys, and cirques are not characteristic of their work; they flowed from centers that were hundreds of miles from their terminals, and hence carried some materials great distances; they carried little or no débris on their surfaces, and as they had no side boundaries they had no lateral moraines. Their morainal material came largely from the underlying bedrock; their melting took place over a wide front, with the result that their terminal moraines and outwash deposits are of vast extent; they overrode hill and vale, and hence much of their deposit is independent of topography.

Continental Glacial Erosion

The Floor. — The continental glaciers were powerful engines of erosion; over wide areas they removed all soil and weathered rock, and plucked and abraded the floor, giving rise to gently rolling surfaces that display glacial striae, groovings, and roches moutonnées on fresh unweathered rock. These roches moutonnées have a smooth, gently sloping up-glacier side and a more abrupt down-glacier slope. The glaciers overrode hills and sculptured them into shapes similar to the smaller roches moutonnées. But toward the southern edge of the ice advance, pronounced erosional features are less conspicuous, although striae and grooves are abundant; in places the weathered surface rock has not been entirely removed. Erosional features in the regions of continental glaciation are, on the whole, less evident than depositional features.

Glacial Lakes. — Throughout the glaciated region of eastern North America are hundreds of thousands of lakes, large and small, that have resulted from glaciation. In the northern areas they have been formed in part by the flooding of rock basins scooped out by the ice. Somewhat farther south in the glaciated area, a more important cause is the formation of depressions by irregular deposition of débris, and especially by the forming of morainal dams across valleys. So numerous and closely spaced are these glacial lakes that what would otherwise be a little known wilderness in northeastern North America is made accessible to travel by canoe. In Minnesota alone there are 10,000 lakes. Of the many beautiful lakes that lend charm to the scenery of the rolling country of New England, the Adirondacks, Minnesota, Canada, and Norway, by far the greatest number have resulted from the work of glaciers. In the unglaciated country of the southern and central United States, lakes are rare or wanting.

Transportation by Continental Glaciers

The continental glaciers carried great quantities of rock débris which later was dropped on the land or spread afar by the action of water. Most of the material of the drift has not been moved more than 50 miles, although some of it has travelled hundreds of miles. This has been determined by the finding of materials far distant from their place of origin. For example, slabs of native copper that came from the great copper lodes of the Keweenawan Peninsula of Michigan have been found as far south as Missouri; and boulders of an unusual conglomerate containing reddish pebbles of jasper found throughout Ohio have been traced to a bed on the north shore of Georgian Bay, Canada. In northern Sweden, glacial boulders of copper ore transported by the ice were traced back in the direction of ice flow and resulted in the discovery, by electrical surveys in 1925, of Sweden's most important copper mine. Of unusual interest in this connection was the finding of several isolated diamonds of good quality in glacial deposits in Wisconsin, Michigan, Ohio, and Indiana. Their position in the deposits indicated clearly that they had been transported and deposited by the ice, but their source is as yet unknown. It is presumed to be somewhere to the northward in Canada.

Deposits of Continental Glaciers

The disappearance of the continental glaciers left the vast regions overridden by them mantled with glacial débris of various kinds and arrangements. These deposits have exerted a profound effect on the social and economic development of the countries with a glacial past, for the fertility of the soil, the drainage, and in part the topography, have been conditioned by such deposits.

As in the case of the existing glaciers previously discussed, deposits were formed either of material dropped directly from the melting ice or of sediments collected by running water. Therefore both glacial till and stratified drift are widespread in the glaciated countries.

The Till or Unstratified Deposits

The Till Sheet. — When the ice melted, its load of débris was dropped as a sheet of till that overspread the land with a variable thickness ranging up to several hundred feet. It consists of old soil, ground-up rock, small fragments, pebbles, and boulders in heterogeneous disorder, and commonly the adjacent boulders have no similarity in composition (Fig. 88). Many of the pebbles and boulders are polished, scratched, and faceted (Fig. 85). Glacial boulders are so common that one may pick them up at random in the glaciated regions; the many stone fences of New England attest their abundance. In places they are so numerous as to render the soil untillable. This bouldery character of the till, together with the abundance of clay in the matrix, has given rise to the term *boulder clay.* Variations in thickness of the till sheet make the surface somewhat irregular, even where the land was originally almost level. In some localities there are smooth, oval-shaped hills of till, elongated in the direction of ice movement. These are *drumlins;* it is not known definitely how they assumed their form and position (Fig. 89). They are especially common in New York, eastern Massachusetts, and eastern Wisconsin. In general, they average about a half-mile in length and 100 to 150 feet in height; and commonly they occur in clusters.

Moraines. — Near the margins of continental glaciation there are belts a few hundred feet to several miles broad that look as though an

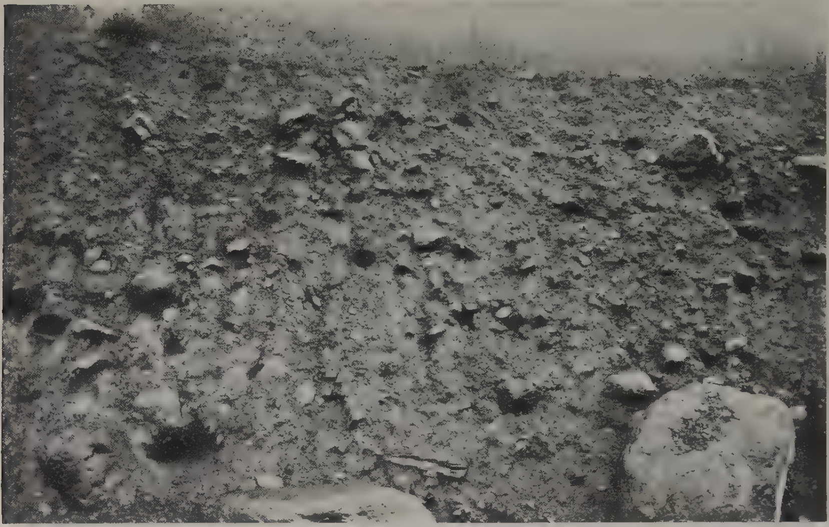


Fig. 88. — Glacial till, Bangor, Pa., consisting of unassorted boulders and fine material. (U. S. Geol. Surv.)

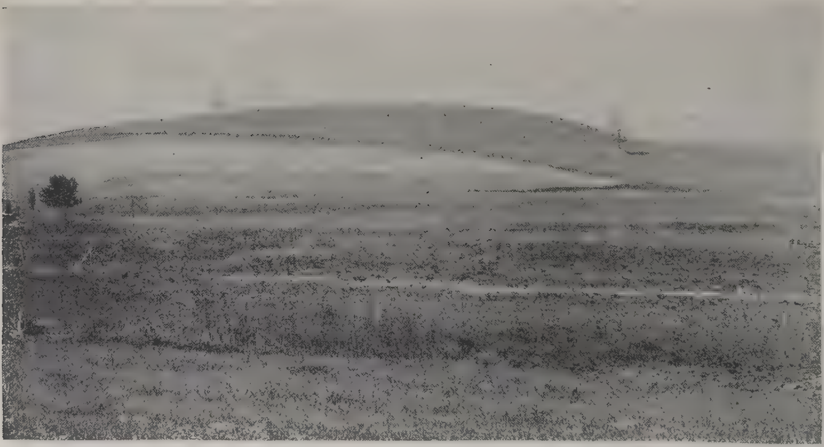


Fig. 89. — Drumlin near Newark, N. Y. (U. S. Geol. Surv.)

army of excavators had dug a maze of depressions and stacked the débris in nearby dump piles. There is bewildering confusion in the haphazard arrangment of pits and mounds. Examination of the material itself reveals glacial pebbles and the lack of sorting characteristic of till (Fig. 88). This is the work of the glacier, and these features are typical of its terminal moraine. In its extensiveness, it bears little resemblance to the small terminal moraines of valley glaciers, for these large ones may be followed almost without break across states and countries. However, the peculiar *knob-and-basin* topography in the great moraines has the same general origin as in the analogous valley moraines; it is due to the irregular deposition of débris liberated from a fluctuating ice front. If the ice, for a period of time, melted about as rapidly as it advanced, much material was dropped in one place. As the ice front oscillated, and projecting lobes and embayments changed their position and shape, the frontal deposition was necessarily hummocky. Some of the knobs rise 100 or 200 feet above the basins, but elsewhere the relief is much less. The depressions commonly are occupied by marshes or ponds that are numbered in the thousands. In some places the terminal moraine consists only of a prominent ridge or a series of separate or overlapping ridges.

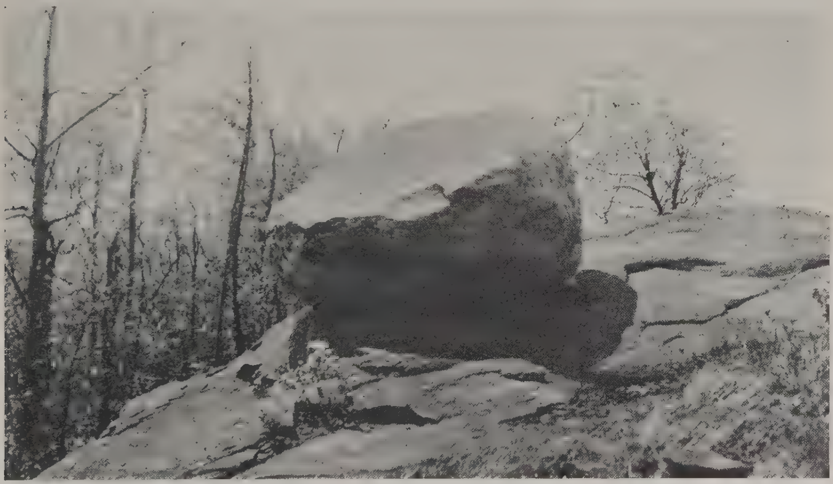


Fig. 90. — Large granite boulder left by the glacial ice on top of a high ridge which is composed of basalt. Mt. Tom, Mass. (Longwell.)

Every considerable halt of the ice front in its retreat gave rise to other moraines, called *recessional moraines*, and the unsteady retreat of the vanishing glaciers can be traced as surely as though they had been under observation. Long lobes of ice must have lagged behind the retreating main sheet, for in many places the terminal and recessional moraines extend parallel to the direction of general movement, indicating deposition along the sides of such tongues. Kettle holes, formed from the melting of residual blocks of ice, are numerous in the vicinity of the terminal moraine.

Erratics. — Unlike the glacial boulders or erratics of mountain glaciers that are found only in valleys, those dropped by continental glaciers occur far and wide over hill and vale. It is not uncommon to see them on the very tops of high hills, in positions they could have reached only through transportation by glacial ice (Fig. 90).

Stratified Drift

The disgorging of torrents laden with sediments, as described for the small living glaciers, must also have taken place on a vastly greater scale from the hundreds of miles of front of a continental glacier. This is not mere supposition, for the extensive deposits of stratified drift that lie within and around the regions of continental glaciation abundantly prove it. However, a greater variety of water-laid deposits emanated from continental glaciers than from valley glaciers, and there is evidence that some were formed under the ice, and others near its edge, in addition to those that were laid down beyond the ice front.

Outwash Plains. — In front of the continental glaciers, valley trains such as those described in connection with existing valley glaciers formed only where well-defined valleys were present. But valleys were infrequent in the open country fronting the continental ice caps, and much of the water's burden was deposited to form great coalescing alluvial sheets, known as *outwash plains* or *frontal aprons*. Like the smaller valley trains, they are composed of water-sorted gravels and sands, built up to considerable depths by streams that rapidly became graded and braided. Some of the finer rock flour supplied the material to make the loess deposits of the Mississippi Valley and in parts of Europe. Southern Long Island was built up as a sandy outwash plain, sloping southward from a large terminal moraine. With the retreat of the ice, outwash plains came to overlie earlier till deposits (Fig. 92).

Some of the outwash plains are prominently pitted by large and small kettle holes, where the outwash materials have slumped after the melting of stagnant ice masses that were partly or entirely buried by the accumulating sediments. Many of these kettles, some as deep as 100 feet, occur in the sandy outwash plain at New Haven, Connecticut.

The sands and gravels of outwash plains are widely used as materials for road and building construction.

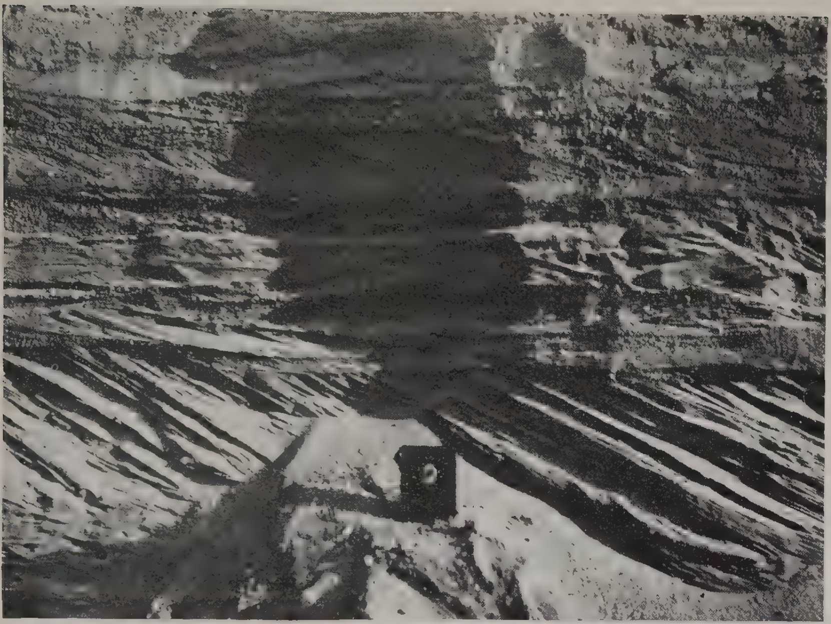


Fig. 91. — Stratified drift. Excavation for Yale Library, New Haven, Conn. (Longwell.)

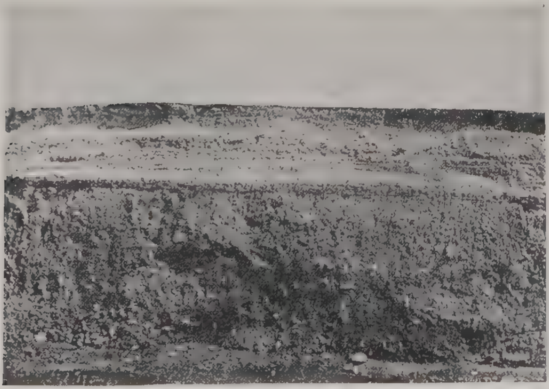


Fig. 92. — Water-laid glacial drift on unassorted till. Columbus, Ohio.

Kames and Eskers. — These are peculiar forms of deposits made by the sediment-laden streams from the melting ice. *Kames* are hummocky hills and ridges composed of stratified drift, which generally occur in clusters with depressions between them. It is believed they were formed at the glacier's edge, where steep cones of water-laid gravel became heaped up against the ice front, and, as the ice front melted, slumped down as rounded hillocks; or under the frontal ice by gravel-laden streams that flowed into depressions or holes; or in shallow basins on top of the ice, from which position the stratified gravels slumped into piles as the ice melted.

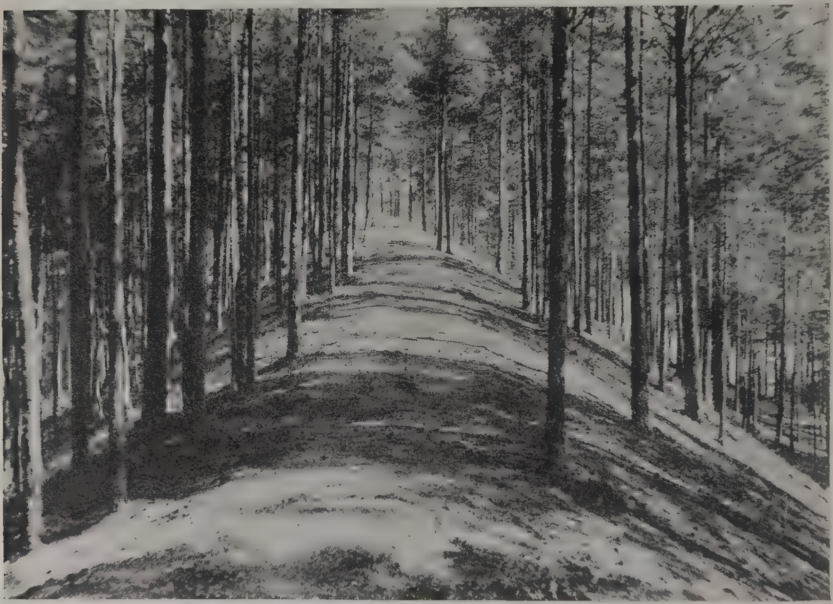


Fig. 93. — The esker of Punkaharju, Puruvesi, Finland. In Scandinavia such a ridge is called an "ose."

Eskers are long, winding ridges of stratified gravel or sand that look like railway embankments. They are only a few feet wide at the top, and range from 10 to 100 feet in height, but some of them are many miles in length. They are striking features of the landscape in Scandinavia, Finland, Ireland, and the northeastern United States (Fig. 93). Presumably they were formed by deposition from streams that flowed in ice tunnels in and beneath stagnant ice. As the stream beds became built up by deposition of gravels, the streams enlarged their tunnels by melting the roofs. Thus there accumulated long, sinuous deposits enclosed by ice. When the ice melted, the gravel and sand slid down to their angle of repose, to form eskers, which trend in the general direction of ice movement.

Lake Deposits. — Remarkably banded clays have been formed in patches within the glaciated regions, notably in Scandinavia and North America. The Swedish geologists call them *varved* clays. In a fresh excavation it is evident that they are horizontal and that most of the individual layers are an inch or less in thickness. But the observer is impressed chiefly with the striking uniformity in thickness of the separate bands and the sharp line of demarcation between layers. If he examines a band closely, he will see that slightly coarser sediment at the bottom grades upward into material so fine that it will not grit between the teeth, and that this alternation is repeated in the individual bands throughout the depth of the deposit. Such fine rhymthmic banding surely could have been formed only in bodies of quiet standing water, such as lakes, and the repeated alternations of coarse and fine material suggest at once a seasonal deposition. And that, apparently, is the manner in which they were formed. The lakes were temporary features, ponded by morainal material or in part by stagnant ice, beyond the melting and retreating ice front. Into them flowed glacial streams carrying sediment. The coarser material was deposited during spring and summer, when the maximum flow of water entered the lake; the finest material, at the top of each varve, represents the slow settling during fall and winter, while the lakes and streams were frozen over most of the time. Thus each layer, grading from coarser to finer material, represents one year's accumulation. Professor De Geer, in Sweden, realizing this, counted the annual layers and reached the conclusion that Stockholm was under ice 9000 years ago; and that it took about 5000 years for the ice to retreat to a point 270 miles northwest of Stockholm.

MARGINAL GLACIAL LAKES

The retreating ice cap had along its front places where the ground sloped toward it, and naturally in such sites water accumulated to form lakes. These disappeared when the impounding ice melted. Lakes of this origin, some large, some small, abounded along the ice fronts; the beaches, outlets, and deposits are still visible. One of the most noteworthy of these, known as *Glacial Lake Agassiz,* covered a large area in North Dakota, Minnesota, and southern Canada. Its extensive deposits constitute the fertile wheat lands of that region. Similarly, a series of lakes of changing outline occupied the area of the

Great Lakes and St. Lawrence Basin and extended across eastern Ontario. Their outlet was for part of the time through the Mississippi River, later through the Mohawk Valley to the Hudson River, and finally through the St. Lawrence.

The Glaciers of the Ice Age

In the light of our knowledge of the behavior and results of existing glaciers, the features just described lead inevitably to the conclusion that extensive ice caps once overspread great continental areas during the Glacial Period. The evidence is now so plain and convincing that it is difficult to realize that the idea was first proposed in 1837 by Louis Agassiz, and that a quarter of a century elapsed before the glacial theory was generally accepted. The older ideas that the débris was transported by the Deluge, or by floating icebergs moved by the Deluge, were dispelled slowly.

Maps showing the directions of the glacial striae, roches moutonnées, eskers, drumlins and terminal moraines prove that the glaciers did not originate at the poles. In North America there were three main centers, from which the ice spread outward in all directions. In Europe the main ice cap centered in Scandinavia and spread into Russia, Germany, Holland, and the North Sea. The thickness of this ice sheet is thought to have been 6000 to 7000 feet in Scandinavia and about 1500 feet in the Harz Mountains. The Labrador sheet was at least 6000 feet thick in northern New England, and may have been thicker farther north.

At the same time that the great ice caps were spreading over the lands, mountain glaciation was also extensive; and while the ice caps of America and Europe disappeared, the mountain glaciers receded to their present-day smaller proportions.

Extended studies of the deposits of the Glacial Period show that there was not a simple advance and retreat of each ice cap. Glacial history is far more complicated. In North America there were at least four, and perhaps five, separate advances and retreats of the ice. These are recognized by later till sheets that cover earlier drift, or that overlie interglacial soils, and by other criteria. There was a widespread retreat of the ice in each interglacial epoch, although it is not certain that the ice disappeared entirely preceding each readvance.

The recession of the ice caps left the lands greatly changed. Not only were they eroded, covered by deposits, and dotted by lakes, but the preglacial drainage systems were profoundly altered. New streams originated, old channels became buried, directions of flow became reversed, and many waterfalls were formed. Deranged drainage on a large scale is evident in the Ohio River basin and in parts of New England.

But what caused these great continental ice caps? The answer to this question has puzzled investigators ever since the fact of continental glaciation became firmly established, but as yet there is no generally accepted explanation. It is becoming clear, however, that they probably did not owe their origin to a single cause but to a combination of causes. The various hypotheses advanced are discussed in Part II of this book.

CHAPTER VI

SUBSURFACE WATER

OCCURRENCE, MOVEMENTS, AND RECOVERY

Water on the surface of the Earth wears away the highlands and fills the depressions with sediment. It is a leveling agency that tends in the long course of time to make the land smoother. But in addition to the work of leveling at the surface, water performs other geological work of great importance, chiefly chemical, beneath the surface.

Source and History of Subsurface Water. — A part of the subsurface water ascends directly from the reservoirs of fluid rock within the Earth; another part is water that was trapped in the sedimentary strata at the time the sediment was deposited; but most of it is that part of the rainfall that sinks into the soil and into the bedrock below. This water, once it has penetrated into the ground, may find its way back to the surface as springs and join the run-off; it may be drawn to the surface by capillary attraction through the pores in the soil and be evaporated; it may be sucked up by plants and be evaporated through their leaves; it may find its way to the sea through underground channels without returning to the surface; it may be held for indefinite periods within the pores of the rocks; or it may form combinations with the molecules of certain minerals and so become fixed in the rocks.

Porosity of Soils and Rocks. — The porosity of a material is its property of containing open spaces or interstices and may be expressed as the proportion of the total volume of the material not occupied by solid matter. This property varies greatly in different materials but the existence of subsurface water depends upon the fact that all rocks and soils are more or less porous.

It has been shown that spheres of equal size placed together so that each one touches all those that surround it leave unfilled 25.95 per cent to 47.64 per cent of the total space that they occupy. The unfilled space is pore space and the difference in amount depends upon the manner in which the spheres are grouped and not at all upon their size. Ordinary sedimentary matter is not composed of perfect spheres but of irregularly shaped, more or less rounded grains of varying size. If these grains are well sorted so that those of one size are collected together to the exclusion of other sizes, the porosity of the material approaches and, because of the irregular shape of the grains, may even exceed the maximum for spherical grains. Poorly sorted sedimentary materials have a much lower porosity because fine grains fill the interstices between the larger ones and reduce the amount of open space.

The porosity of materials of the same general kind varies considerably but the following table gives a good idea of average amounts. These figures show that soils have the highest porosity in spite of the poor sorting of their constituents. This is explained by the poorly compacted condition of most soils. They are kept loose and aërated by worms, ants, burrowing animals, and vegetation as well as by plowing and agriculture in general.

Unconsolidated Material		Consolidated Material (Sedimentary Rocks)		Igneous and Metamorphic Rocks
Sand and gravel	35 per cent	Sandstone	15 per cent	Average porosity less than one per cent
Clay	45 "	Shale	4 "	
Soil	55 "	Limestone	5 "	

Igneous, metamorphic, and well-consolidated sedimentary rocks commonly have a porosity much higher than that given because systems of fractures develop within them. Limestone may be rendered very porous by the development of cavities due to solution. Figure 94 and the accompanying description illustrates the relation between rock texture and porosity.

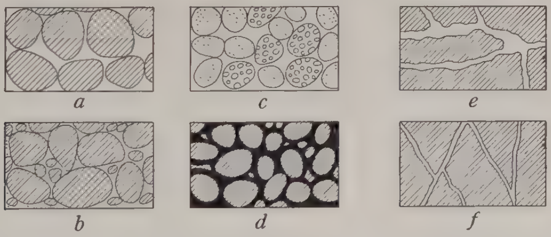


Fig. 94. — To show relation of rock texture to porosity. *a*, well-assorted sedimentary deposit having high porosity; *b*, poorly-assorted sedimentary deposit having low porosity; *c*, well-assorted sedimentary deposit made of pebbles which are themselves porous, giving the deposit as a whole very high porosity; *d*, well-assorted sedimentary deposit whose porosity has been diminished by deposition of mineral matter in the interstices; *e*, rock rendered porous by solution; *f*, rock rendered porous by fracturing. (U. S. Geol. Surv.)

Aquifer. — The term *aquifer* is applied to a layer or other body of loose sediment or consolidated rock that yields water in considerable quantities.

Forces that Control Water in Rocks. — Gravity and molecular attraction are the two forces that control the movement of subsurface water. Gravity causes the water to percolate to lower levels or to move laterally down a grade. It also causes the water to issue from springs and flowing wells, forced out by the weight of the water behind. If all the interstices in rocks were large, gravity would be the only controlling force. But, since the actual interstices in many rocks are minute, the molecular attraction of the rock substance acts across the openings and holds the molecules of water firmly in place in spite of their tendency to move downward. This force is called *adhesion.*

Water-Yielding Capacity of Rocks. — Not all of the water contained in the pore space of a rock will be yielded to the natural drainage or to such artificial openings as wells. It has been shown that the size of the grain of a sediment does not affect its porosity but the last paragraph explains how that size does determine the percentage of the contained water that the sediment will yield. A porous and thoroughly saturated clay may be *impervious* to water at usual pressures because the water already in it is held firmly by the molecular attraction of the particles of clay. On the other hand a compact, granular rock such as a granite, with a few scattered fractures, is much less porous but yields the water that it does contain more readily. It is more *pervious* than the clay.

The Situation of Subsurface Water

The water in the ground fills all the pore space up to a certain definite level called the *water table.* The zone below the water table is known as the *zone of saturation* and the water contained in it is called *ground water.* The region between the water table and the surface of the ground is called the *zone of aëration* and the water in it *suspended subsurface water* (Fig. 95). The water table is not a level surface but rather a subdued replica of the land surface beneath which it lies. It is arched up under the hills, sinks a little below the valleys and intersects the surface in some low-lying tracts to form lakes, swamps, or springs.

The level at which the water table stands in any particular region depends on the rainfall and the topography. It rises or falls in response to wet or dry seasons and stands within a few feet of the surface in humid regions though it may be several hundred feet in depth in those that are perennially dry. If climatic conditions are the same the water table will stand farther below the hill tops in a rugged region than in one less rugged because the deeper valleys depress the water table near them and give the whole a greater gradient.

The Zone of Aëration. — The thickness of the zone of aëration varies directly with the depth of the water table. It may be nonexistent or it may be several hundred feet thick, though ordinarily it varies between a few feet and a few tens of feet. It is the zone in which suspended water, present in small quantities only, is held in place despite the downward pull of gravity. Wherever soil is present, some water descending from the surface is retained by minute interstices and is thus available for plant growth. In addition, just above the water table the small openings contain water drawn up from the zone of saturation by capillarity. The presence or absence of an intermediate zone between this *capillary fringe* and the *soil water zone* depends on the depth of the water table.

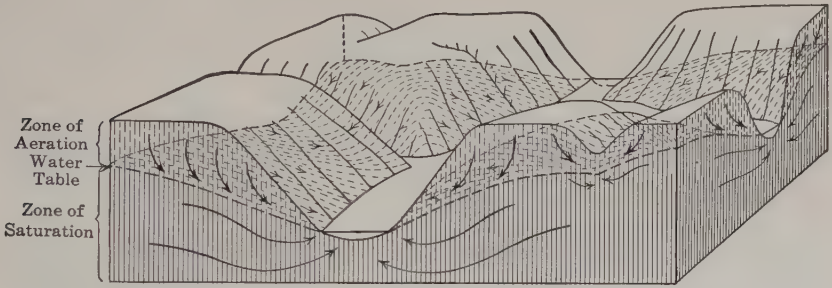


Fig. 95. — Showing the relation of the water table to the surface of the ground; the directions of movement of the subsurface water (indicated by arrows); and the zones of aëration and saturation.

For the sake of simplification the region is regarded as underlain by a single stratum of uniform porosity.

The Lower Limit of the Subsurface Water. — Deep down within the crust of the Earth the weight of overlying matter exceeds the crushing strength of rocks. At such depths the interstices must cease to exist and consequently subsurface water is excluded. This theoretical depth is impossible to state accurately but it lies a number of miles beneath the surface. Actually ground water becomes very scarce after a few thousand feet of rock have been penetrated. Many mines are dry 2000 or 3000 feet down except where deep fissures are encountered. Some aquifers have been encountered at depths as great as 6000 feet, but as a rule wells drilled more than 2000 feet produce little water.

Ground water must be regarded then as a great body of water filling the interstices in all soils, loosely consolidated sediments, and solid rocks between a sharply defined upper boundary called the water table and an indefinite lower boundary ranging from a small depth to many thousand feet below the surface. Lower than this even the most persistent fissures must be closed and dry.

Perched Water Tables and Impervious Beds Lying Below the Water Table. — The conditions outlined above are not always completely fulfilled. In regions where the main water table lies at a considerable depth below the surface an impervious stratum in the zone of aëration may hold a local zone of saturation suspended far above the main one. This zone has its own water table called a *perched water table* (Fig. 96). There are also many impervious strata that extend far down into the zone of saturation. A well that penetrated these strata would be dry even after it had passed some distance below the general level of the water table in the surrounding region. Many such impervious strata would cause the water table to be ill-defined.

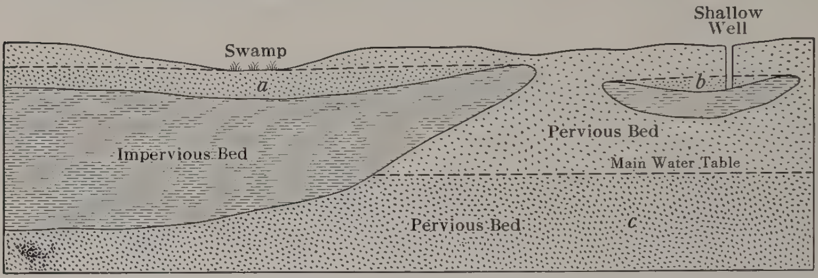


Fig. 96. — Showing perched water tables caused by impervious beds interstratified with pervious beds, and the depression of the main water table by an impervious bed that extends down into the main zone of saturation; *a* and *b* are perched bodies of ground water, *c* is the main zone of saturation. The difference in density of the dots represents different amounts of contained water. (Adapted from Gregory.)

Return of Ground Water to the Surface

Motions of the Subsurface Water. — Above the water table the movement of subsurface water is directly downward (Fig. 95). Below that level the ground water proper moves in the same general direction as the surface run-off but more slowly, because the friction caused by its passage through the interstices of the rock retards its motion. The continuous addition of parts of this slowly moving sea of ground water to streams and lakes and its seepage at the surface or concentration into springs is the cause of the continuous flow of streams. Were it not for the restraining influence of friction the subsurface flow would be as rapid as that at the surface and every rain would mean a flood and every stream would run dry between storms.

The more deeply the ground water penetrates beneath the surface in uniform material the more resistance it meets due to friction and the more slowly it moves until, at the greatest depths to which it can go, it is stagnant, held motionless by the molecular force of the rocks.

Hillside Springs. — Wherever the surface of the land intercepts the water table, ground water emerges in some fashion. When the water oozes out along the line of contact a seepage results, but if the circumstances are such that water is concentrated in quantity sufficient to form a distinct current, an ordinary spring is formed. This is illustrated in Fig. 97. Figure 98 shows the wall of a canyon that has cut the water table so that the water issues in a series of springs.

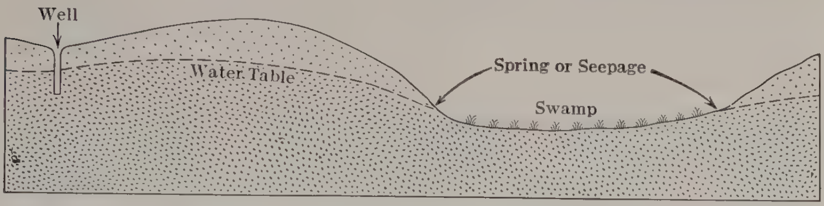


Fig. 97. — Showing conditions favorable for wells, hillside springs, and a swamp or lake.

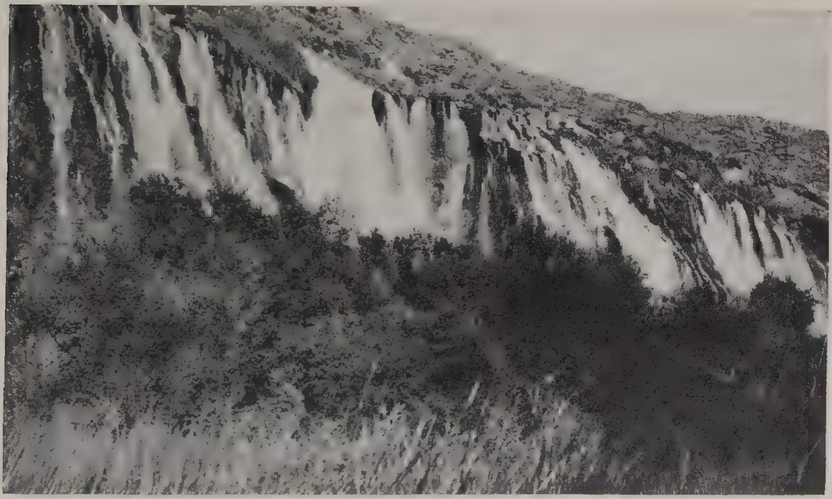


Fig. 98. — Thousand Springs, Snake River canyon, Idaho. The wall of the canyon cuts the water table and allows the ground water to issue as a series of springs. (U. S. Geol. Surv.)

Wells. — An ordinary well is an opening dug or drilled into the ground to a depth sufficient to penetrate the saturated zone (Fig. 97). The water percolates into the opening from the saturated material around it but does not rise in the well above the level of the water table.

Fissure Springs. — When surface water enters and saturates a pervious, inclined layer confined between two impervious layers the weight of the water in the aquifer generates hydrostatic pressure. If the aquifer is cut by a fissure that leads to the surface, the water will be driven up along the fissure and issue as a spring or a series of springs lined up along the fissure. Such springs are often called fissure springs. Figure 99 illustrates the following necessary conditions: a catchment area where water may enter the aquifer; a pervious stratum to serve as aquifer lying between two impervious strata; an inclination of the strata sufficient to cause the water in the pore space to exert pressure on that down slope from it; a fissure to cut off the normal course of the water and furnish a channel to the surface; a point at which the water may issue below the level at which it entered the aquifer. The difference in elevation between these two levels is known as the *hydraulic head* which, together with the size of the pore space, determines the force with which the water will reach the surface. There is always a considerable loss of head due to friction.

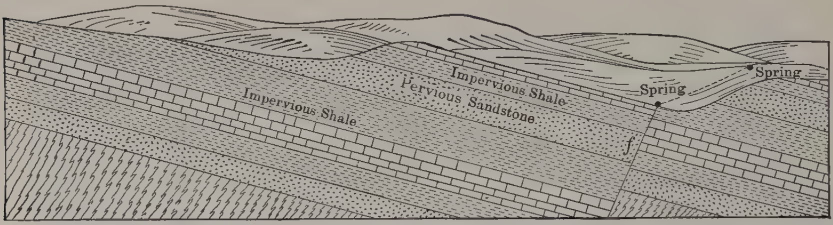


Fig. 99. — Illustrating conditions favorable for springs, if fissures, such as *f*, are present.

With such an arrangement as is postulated in Fig. 99, springs might be expected at various points along the line of intersection of the fissure with the surface, wherever suitable channels for the upward flow of the water exist. Such springs are usually very steady in their flow and are less affected by droughts than ordinary hillside springs. They are usually cold, but the water may come in contact with heated rocks and issue as warm springs. This is probably the explanation of the warm springs that occur at various places in the Appalachians, as at Hot Springs, Virginia. The water sometimes dissolves unusual amounts of mineral matter and gives rise to *mineral springs*, as at Saratoga, New York; Carlsbad, Bohemia; Bath, England; Wiesbaden, Germany; and Vichy, France. These springs contain considerable though different amounts of sodium, chlorine,

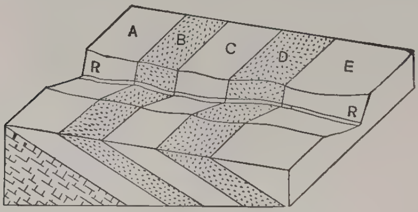


Fig. 100. — To illustrate entrance of water into pervious rock layers, or strata. *ACE*, impervious beds; *BD*, pervious beds; *RR*, course of river.

carbon dioxide, sulphate, and smaller quantities of calcium, magnesium, and many other elements.

The condition under which pervious beds become filled with water is important, not only for fissure springs, but also for artesian wells, described below. The pervious layers *BD*, Fig. 100, become filled, not alone by the rain that falls on their exposed surfaces, and by the water that is shed upon them from the higher impervious slopes *A* and *C*, but from the river water that is concentrated from the watershed above.

Artesian Wells. — If under conditions similar to those described above, but in the absence of a fissure, a hole is bored down to the aquifer, the water will rise above the zone of saturation and produce an *artesian well.* The height to which the water rises above the zone of saturation depends upon the pressure, which in turn depends on the height of the water column, or " head " in the aquifer above its upper surface where the well is drilled; the size of the pore space of the aquifer, which determines the loss of head due to friction; and perhaps to some extent on the weight of the overlying formations which tends to compress the water-bearing stratum and force the water out. In many cases the water rises to the surface or above it. The alternating pervious and impervious beds may have the form of a basin (Fig. 101, *A*) but that is not a necessary condition. The arrangement illustrated in Fig. 101, *B* is just as suitable; the water rises through the artificial opening because the diameter of the well is greater than that of the openings that constitute the pore space of the aquifer and the water encounters less resistance than by pursuing its course underground. The water spouts from many artesian wells as in a fountain (Fig. 102).

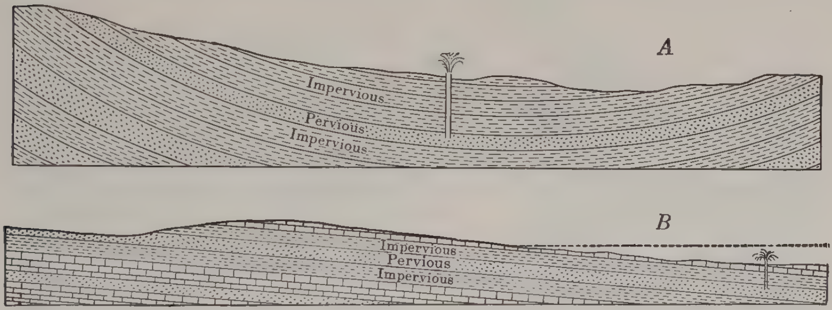


Fig. 101. — Sections showing conditions favorable for artesian wells. Vertical scale exaggerated.

A. Strata are folded so as to form a basin.
B. Strata are inclined in one direction only.

Artesian wells cannot be made simply by boring deeply unless the requisite geologic conditions are present. Deep wells bored into rock so as to intercept the water table are often called artesian wells but this is an incorrect use of the term; there is no difference in principle between wells of this kind and ordinary shallow dug wells. Some of the most important water-bearing formations in the United States which furnish artesian wells are the Dakota sandstone, which comes to the surface along the Rocky Mountains and Black Hills, and underlies large parts of North and South Dakota, Kansas, and Nebraska, and extends into Canada; and beds of sand that underlie the Atlantic Coastal Plain from Long Island to Texas. The conditions are generally unfavorable for artesian wells in the uplands of New England because the veneer of glacial till and lake and stream deposits is too thin and discontinuous to form artesian aquifers, and it overlies jointed and faulted metamorphic or igneous rocks in which the openings are too closely spaced or the individual fractures are not sufficiently large to establish artesian circulation.

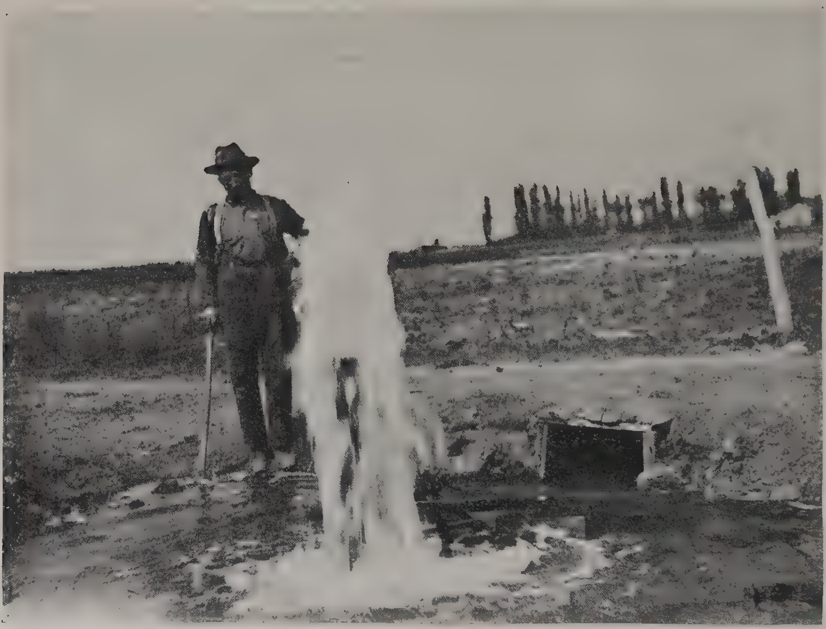


Fig. 102. — Artesian well, near Provo, Utah. (U. S. Geol. Surv.)

The depth to which wells must be bored before artesian water is attained is very great in some places. In Berlin, St. Louis, and Pittsburgh the necessary depth is about 4000 feet, and depths of 1000 feet are not uncommon. Along the Atlantic coast, on the other hand, artesian wells are generally shallow — from 100 to 300 feet. The volume of water may be very large; the great 12-inch well of St. Augustine, Florida, with a depth of 1400 feet, supplies 10,000,000 gallons a day. Where many wells are put down close together the drainage basins are likely to interfere with each other, and the withdrawal of water lowers the pressure to such an extent that the water will no longer rise above the level of the aquifer.

GEOLOGIC WORK OF SUBSURFACE WATER

Water underground is an important geologic agent. Its chief work is to *take substances into solution, carry them elsewhere, and perhaps redeposit them;* this work is, therefore, largely chemical in its nature. In the discussion of the decay of rocks and the formation of soil it was shown that certain constituents such as the alkalies in the rock-forming feldspars go into solution and are removed; and that calcium carbonate, a common rock-making material, is dissolved and carried away by water containing carbon dioxide. These actions are accomplished by atmospheric water as it passes underground, and may thus be regarded as the first stages of the work of subsurface water. Rivers carry a large part of their burden in solution, and ultimately deliver this material, dissolved by subsurface water, to the sea. Finally the formation of salt lakes, and the deposits that occur in them, illustrate the work of solution, transportation, and deposition.

These facts illustrate the general chemical work of water, partly on the surface and partly underground, but there are certain features that demand particular consideration.

Solution. — The solvent action of rain water passing into the soil and rocks is greatly increased by the substances that it carries with it, or that it may otherwise obtain. In its passage through the air it dissolves carbon dioxide and oxygen, together with minute amounts of other materials, and is thus equipped for doing chemical work (Fig. 103). In passing through the humus and upper soil of humid regions it may absorb much more carbon dioxide as well as organic acids produced by the decomposition of organic matter. In many places, particularly in volcanic regions, volatile substances, especially carbon dioxide, are evolved from the depths, and may dissolve in the subsurface water, thus augmenting the quantity of chemical reagents present in it. In addition, as the water passes into deeper zones the pressure increases and it may come in contact with heated rocks and have its temperature raised, both of which changes greatly increase its chemical efficiency. The amount of gas, such as carbon dioxide, which water can hold in solution, is directly proportional to the pressure. The amount decreases as the temperature of the water rises but this is more than compensated by the increased chemical activity of hot water.

Pure water dissolves many substances, but with its chemical efficiency heightened as just described, it attacks the minerals composing the rocks and soils with added energy. It takes some of them, such as gypsum ($CaSO_4.2\,H_2O$), directly into solution. In other cases a chemical reaction takes place and new compounds are formed, some of which are soluble and are carried away, while the insoluble material remains.

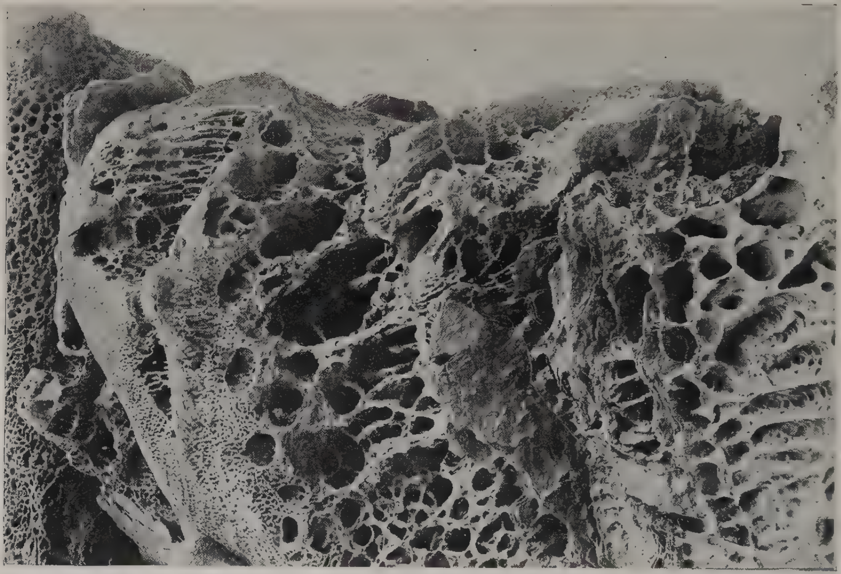


Fig. 103. — Rock whose more soluble parts are being dissolved by the action of atmospheric waters. Wind aids the rain in removing the loosened material. Near Livingston, Mont. (U. S. Geol. Surv.)

The material taken up and held in solution pursues one of two courses, depending on what happens to the water that contains it. It works down into the rocks and is deposited there, or it emerges into the surface drainage and is carried into the ocean.

The process by which the land surface is wasted by solution is known as *chemical denudation,* to distinguish it from the mechanical wear of ordinary erosion. The amount of material so removed each year is very great. Dole and Stabler by assembling a large number of analyses of the waters of the Mississippi, which analyses give the average percentage of the salts it contains, have calculated that 108 metric tons of matter are removed each year in solution for every square mile that it drains.

Results of Solution of Carbonate Rocks. — Aside from the process of soil formation, in which solution plays an important part, the most obvious results of solvent action are seen in the effects it has upon rocks wholly, or partly, composed of carbonates. The most important rock-forming carbonates are those of calcium, magnesium, and ferrous iron; $CaCO_3$, $MgCO_3$, and $FeCO_3$. The carbonate of calcium especially underlies vast stretches of land, as beds of limestone hundreds or even thousands of feet thick. Besides this, beds of sandstone commonly contain a cement of calcium carbonate that binds the grains of sand together. Since these carbonates, especially calcium carbonate, are attacked by water containing carbon dioxide, many such rock masses must be continually dissolving and wasting away. This is suggested by the fact that in those places where limestone, or calcareous sandstone, is the bedrock the water is always *hard*, i.e., contains lime in solution.

Sinks and Caverns. — In regions where limestone forms the bedrock the surface water works down through joints and fissures and enlarges them by solution. When the water reaches an insoluble layer, such as one of clay or shale, it is stopped in its descent and spreads laterally, finding its way through the rock fissures along the natural drainage slope. These fissures are also enlarged by solution until they become distinct water channels. As the latter enlarge they form *caverns*, while the holes or pipes, leading down to them from the surface above, are termed *sinks* (Fig. 104). Sinks are also formed by the collapse of the roof, weakened by solution, into the cavern beneath.

Some of the individual chambers hollowed out in the rock are 100 feet or more high, and several hundred feet broad. They are connected by intricate passages. The floor on the insoluble stratum may be quite level for long distances though it is broken through in many places when the water excavates new passages and chambers at a lower level (Fig. 105). The final result may be several sets of such rooms and galleries, one above the other. Some well known natural bridges have resulted from partial collapse of cavern roofs.

The limestone regions of the Middle West and the South are noted for their caverns, some of the best known being Mammoth Cave in Kentucky, 10 miles or more long, with 30 miles of winding passages; Wyandotte Cave in Indiana, Luray Caverns in Virginia, and the Carlsbad Cavern, New Mexico. In some places the rocks are honeycombed with passages and almost the entire drainage may pass underground. Large rivers disappear from sight, and after a devious journey below come to the surface again in a different drainage area. Some give rise to huge springs, such as Silver Spring in Florida which has a flow so large that the resulting stream is navigable for small steamers. Other underground streams are forced up as great springs in the sea, not far from land.

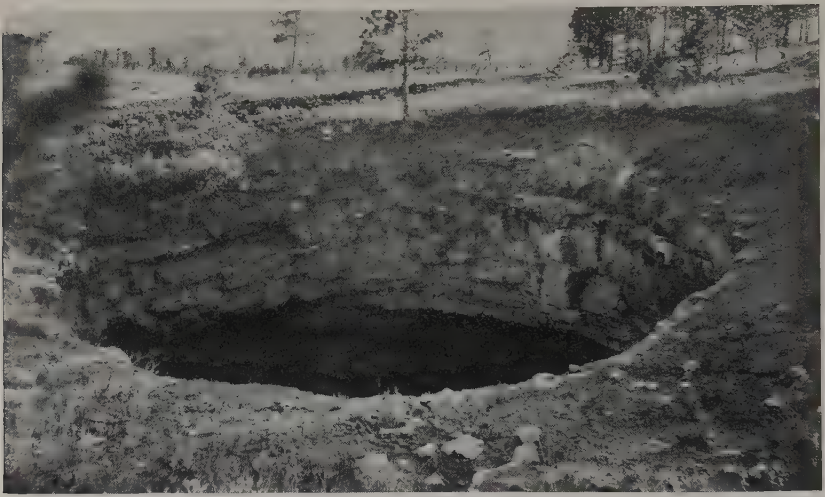


Fig. 104. — Opening to sink in limestone beds; near Cambria, Wyo. (U. S. Geol. Surv.)

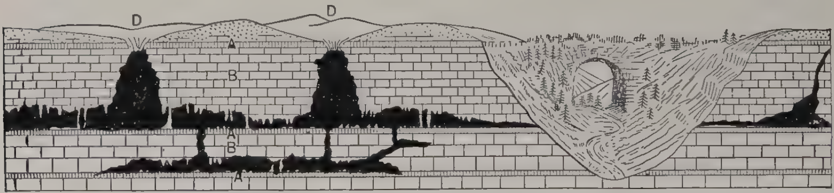


Fig. 105. — Illustrating the formation of caverns and sinks in limestone. *AA*, clay beds; *BB*, limestones. The arch is the remnant of the roof of a former cave, forming a natural bridge. *DD*, sinks, leading to caverns below. (Modified from Shaler.)

Deposition and Cementation. — From what has been previously stated it is clear that there is an upper belt in the Earth's crust where mechanical and chemical changes and destruction are going on. It is known as the *zone of weathering*, and extends downward to the level of the water table. Material is being constantly leached out of this zone and carried down in solution into the ground water. This matter is either carried away by the drainage, or deposited in the pores and other cavities in the rocks. The lower limit to which this can proceed is uncertain. It appears to depend on several conditions and it probably varies in different places. Thus the intervention of impervious rock layers, or the charging of the rock pores with gas under pressure, would hinder, or perhaps prevent, further downward movement in a given area. It is in this belt, whatever its thickness, that the sediments are being solidified and cemented by the silica, calcium carbonate (calcite), and other substances deposited in them, and it is, therefore, known as the *zone of cementation.*

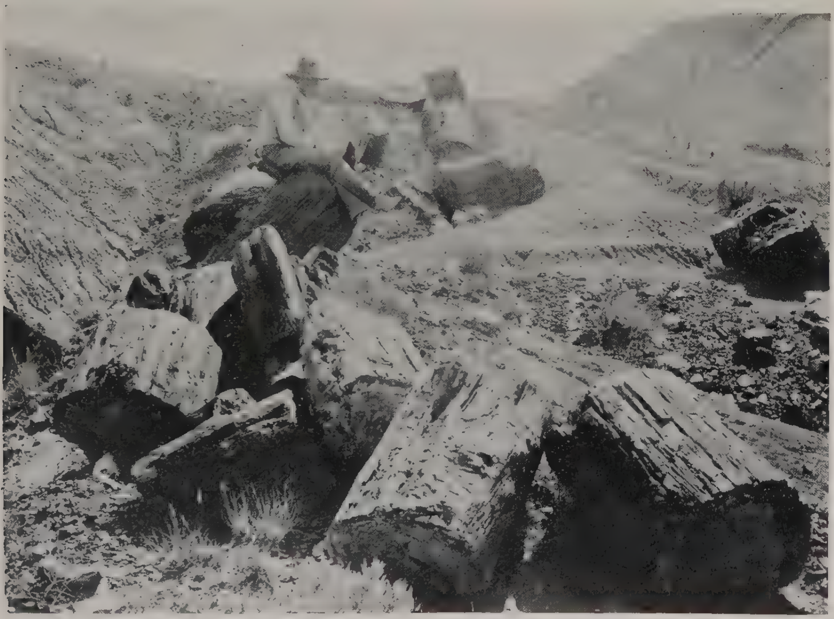


Fig. 106. — Petrified log sections. Petrified Forest, Arizona. (U. S. Geol. Surv.)

There are several reasons why much of the material dissolved by the ground water is redeposited by the same agency within the pores of the rocks. The most important of these are (1) Evaporation of water in open spaces such as caves. (2) Loss of dissolved gas such as CO_2 by warming or evaporation. (3) Cooling of the water. (4) Lowering of the pressure. (5) Chemical reaction between the solution and the rock through which it is percolating.

Replacement. — The deposition of material from solution in ground water does not always take place in openings already present. Frequently the water will dissolve a certain substance and leave behind an equal volume of another substance that it held in solution. Thus a buried tree trunk may be slowly converted into stone (*petrified*) by the solution of its vegetable matter and the deposition of opaline silica (Fig. 106). The fact that such replacements take place volume for volume is shown by the complete preservation of the fine texture of the original material regardless of the relative size of the chemical molecules involved.

Concretions. — The capacity of ground water to dissolve and redeposit is nowhere better illustrated than by the formation of concretions. Concretions are regular, rounded, or strangely shaped nodules that occur in sedimentary strata. They are due to deposition from the ground water which has dissolved the substance from other parts of the same rock or brought in some foreign matter. They are explained in more detail in the chapter on sedimentary rocks.

Geodes. — Water standing or slowly circulating in rounded cavities below the water table deposits a lining of mineral on the walls. Crystals grow from the walls into the mineral-bearing water, and in time there is developed a spherical shell lined with beautiful crystals, known as a *geode*. Many of these features are encountered in mining operations; and some that have been released from the surface rocks by weathering resemble rounded boulders until they are broken open to reveal their true character.

Stylolites. — Irregular, sutured partings characterized by interlocking columns or teeth from a fraction of an inch to several inches in length commonly penetrate soluble rocks such as limestone, dolomite, or marble, essentially parallel to the bedding planes. These partings are known as *stylolites*. They are formed by unequal solution along the opposite surfaces of a fracture or a clay seam. They are localized by original irregularities of that surface which allow the weight of the overlying rock to press more at one point than at another and by the varying solubility of different parts of the rock adjacent to the fracture or seam. Once started, the rock opposite the end of each column is dissolved away more rapidly than elsewhere and the teeth penetrate further into the opposite bed and develop striae along their sides. The insoluble material finally collects as a clay cap at the end of each column.

Some stylolites develop in relatively insoluble rock, such as quartzite. Wherever they occur they indicate the removal of a considerable quantity of rock in solution.

Relation of Subsurface Water to Ore Deposits. — Certain deposits of lead and zinc ores seem to be due directly to the concentrating action of subsurface water that dissolves mineral matter over large areas and concentrates it in a relatively small zone. Still other deposits — mainly iron ores — owe their value to the removal of worthless mineral matter in solution and the consequent increase in the percentage of the iron minerals in a given area. The value of many important silver and copper veins is due to the concentration of the ore minerals from a large to a relatively small part of the vein by ground water percolating downward, thus raising the metal content within the narrow zone to a profitable amount. This process is known as *secondary enrichment.*

Deposits in Caves. — The same process that forms caverns also tends to fill them up. For, after they have been opened by solution, the surface waters seeping down through the rock layers which form the roofs dissolve more calcium carbonate, and deposit it in the caverns, producing stalactites, stalagmites, and columns. The manner of their formation is as follows: a drop of water, charged with calcium carbonate, leaking through to the roof hangs there for a time. While at rest it evaporates a little, loses some carbon dioxide, and consequently deposits some calcium carbonate. Finally, as more water is added from above, it is forced to drop, and falling on the floor below it evaporates still more and leaves another deposit. Thus long, pendant, icicle-like incrustations called *stalactites* grow downward, and broader, dome-shaped

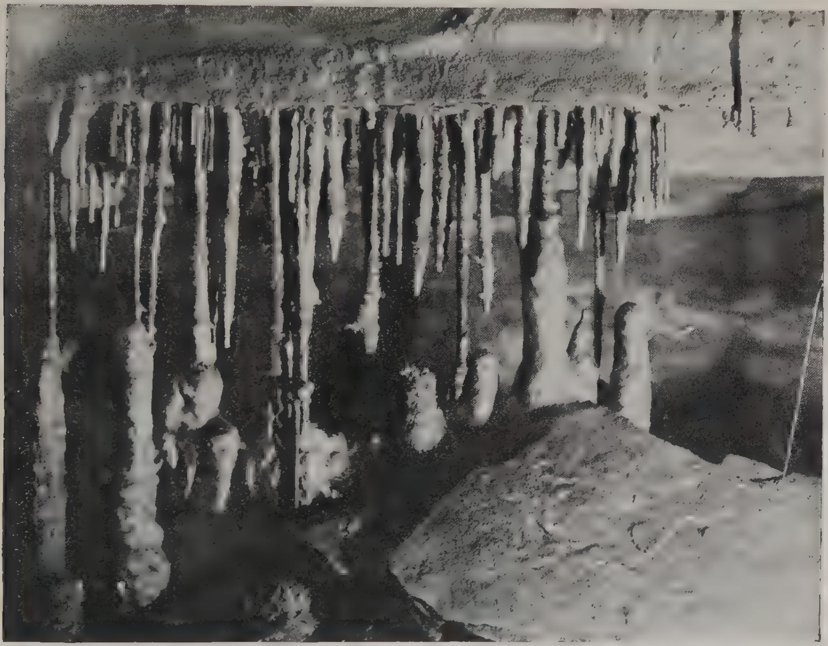


Fig. 107. — Stalactites, passing below into stalagmites, along a roof-crack. Marengo Cave, Indiana. (U. S. Mat. Nus.)

masses called *stalagmites* rise up from the floor immediately below. Some of these finally unite and produce columns. These forms are most likely to develop along fissures in the roof of the cave (Fig. 107).

In past times caves have served as refuges for primitive men who inhabited them, or as dens for wild animals. Because of this the bones of men and animals, stone implements, and other objects have accumulated in them and been sealed up, in the deposits of calcium carbonate on their floors. Relics of this kind, especially in certain parts of Europe, have revealed much concerning the life and degree of culture existing in prehistoric time.

Deposits of Calcium Carbonate by Springs. — The material in solution which is not deposited in the rocks is carried away by the drainage. It sometimes happens that on its way to the sea it comes to the surface again and is temporarily deposited. Springs that deposit calcium carbonate furnish a good illustration of this. Many springs, especially deep or fissure springs, contain much carbon dioxide gas, under considerable pressure. When this water passes through beds of limestone on its upward journey large quantities of calcium carbonate are taken into solution. On arriving at the surface, partly because of evaporation and partly because of loss of gas through the relief of pres-

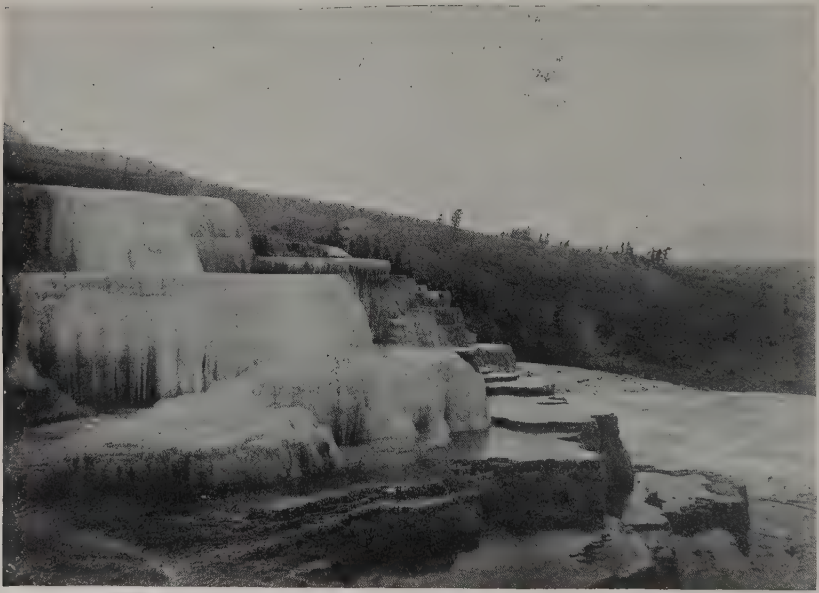


Fig. 108. — One of the terrace formations of the Mammoth Hotsprings, Yellowstone Park, Wyo.

sure, the calcium carbonate is deposited and built up into mounds and terraces, some of which are very beautiful. They are illustrated in the basins and terraces of the Mammoth Hotsprings in the Yellowstone Park (Fig. 108). Other examples of such springs are found in Virginia, Colorado, Banff in Alberta, Carlsbad in Bohemia, Tuscany, and in many other places. Some spring waters contain other mineral substances together with calcium carbonate, or even to its entire exclusion. Many of these springs are used medicinally, as at Saratoga and other health resorts.

In some springs that come from great depth the issuing water is warm, or even hot. This is likely to be the case when the springs occur in regions of active or recently extinct volcanoes such as that in which the Mammoth Hotsprings are situated. In such warm waters the deposit of calcium carbonate may be much increased by the action of primitive forms of vegetable life, the algae, which secrete this substance from the water. In many places the deposition takes place so rapidly that articles suspended in the water become covered in a few days with a coating of the mineral. It is probable that the warmth and chemical activity of the waters of some springs, particularly hot springs in volcanic regions, are greatly increased by gases and vapors that rise from molten magma or masses of hot rock below.

Other Deposits by Springs: Iron Oxides, Silica, etc. — Substances other than travertine may be deposited when underground waters issue at the surface. One of these is the hydrated oxide of iron, or, under certain circumstances, iron carbonate. This is a matter of importance because some extensive beds of iron ore have been formed in this way.

Alkaline Deposits — The soluble substances that are formed by the decay of the rocks in humid regions are quickly washed out of the soil and are carried by the drainage to the sea. In semiarid and desert regions where the rainfall is scanty there is not sufficient water to perform this function and the salts remain in the soil. At times of rainfall they go into solution, and during the subsequent periods of drought, when the water draws to the surface, they are left behind by its evaporation and form the white incrustation on the soil known as *alkali.* This is a common feature in many parts of the western United States.

The common salts that compose alkali are sodium sulphate, sodium chloride, and sodium carbonate. The name alkali is due to the alkaline reaction and taste of the latter. Magnesium sulphate, and calcium sulphate, or gypsum, are aften present as well. These salts are not always furnished directly by rock decay; they may have been originally present in beds of sediments laid down in the sea. Their concentration in such arid regions, with inland drainages, gives rise to salt and alkaline lakes. The irrigation of alkali lands, especially if water is too freely or carelessly used, may bring the salts to the surface in such quantities as to injure, or even ruin the land for agriculture.

Mechanical Work of Water Underground; Landslides. — Subsurface water is of little importance as a mechanical agent. It is conceivable that streams running in subterranean channels may erode and transport, but the circumstances that would permit this are exceptional. A more important mechanical function of subsurface water is its aid in causing landslides, both by helping to overcome the friction of masses of rock, soil, and débris lying on steep slopes, and by adding weight to the mass. The saturated masses of soil and rock act like a semifluid substance and, once started from their insecure foundations at times of heavy rainfall or when loosened by earthquake shock, rush down into the valleys causing great damage and considerable changes in the topography. In high mountains such landslides may precipitate huge trains of broken rock, or talus, down the valley sides, giving rise to rock streams.

CHAPTER VII

LAKES AND SWAMPS

Inland bodies of standing water are called lakes; expanded portions of rivers are also referred to as lakes. The great majority of lakes stand above sea level, but some, including all coastal lagoons, are at sea level; a few, like the Salton Sea and the Dead Sea, are even below. Lakes occur in all parts of the world but since a great proportion of them are a direct result of glaciation, there are more lakes in high latitudes and in high altitudes than elsewhere. There would be no lakes if the surface of the lands were everywhere graded, for then drainage to the sea would be perfect. Where drainage is obstructed and the surface is not at grade lakes occur. Since all the streams are at work destroying obstructions and filling depressions in the ceaseless effort to bring the lands to grade, it follows that lakes are necessarily ephemeral and that all must sooner or later disappear.

ORIGIN OF LAKE BASINS

Warping and Faulting. — Lake basins formed by broad downwarping of the rocks have existed at many times during the Earth's history. We can be sure of this because their shore and bottom deposits are preserved long after the waters that stood above them have been drained away.

Great lake-bearing depressions caused by fracturing and downfaulting of the crust are more common than those caused by folding. The 4000-mile chain of valleys and lakes that includes Jordan River, the Dead Sea, the Red Sea, the upper Nile, and the African lakes, such as Tanganyika and Nyassa, was formed by the sinking down of narrow blocks of the crust, between high steep walls. The formation of some structural basins has occurred within human history in connection with earthquakes. During the New Madrid earthquake of 1811 the surface of the ground was so changed that several new lakes came into existence in the Tennessee portion of the Mississippi flood plain.

Crater Lakes. — One of the most remarkable natural features in North America is Crater Lake in southwestern Oregon, occupying a crater more than 5 miles in diameter in the summit of the Cascade

Range. The arrangement of the rocks which form the crater indicates that this was once a high volcanic mountain, but that the top collapsed and sank to great depth (see page 220). This *caldera* forms a lake 2000 feet deep, without tributaries and entirely dependent on rainfall. The water escapes by evaporation and by seepage into the rocks of the crater rim, reappearing in part as springs at lower levels. Lakes in ordinary craters are found in most volcanic regions. Notable among them are Lago de Bolsena and others in the great volcanic " campagna " surrounding Rome, Lac Pavin in the volcanic plateau of central France, and the " maaren " of the Eifel in northwestern Germany.

Natural Dams across Valleys. — Natural dams may be thrown across valleys in many ways. Chippewa River, which enters the Mississippi about 60 miles below St. Paul, has deposited sufficient débris athwart the course of the larger stream to restrict flow and the Mississippi has in consequence been ponded for more than 20 miles upstream. Landslides occasionally interrupt the flow of streams, converting them into lakes. Several such lakes are known in the region of the Alps, and a new one was formed at the base of the Gros Ventre Range, in western Wyoming, in 1925. Lava flows occurring in regions where streams have already cut narrow valleys are likely to dam the valleys and form lake basins. The lava barriers are eroded with difficulty because made of solid rock. Lac d'Aydat in central France is of this type, as are several of the lakes surrounding the volcanic cones of Mt. Hood and Mt. St. Helens in Oregon and Washington. Again, where valleys are drowned and the force of their stream currents is dissipated by coastal submergence, it is easy for waves and currents to build bars and barriers across their mouths and so convert them into coastal lagoons. Long coalescent chains of these lagoons are found along the Atlantic coast of the United States from Sandy Hook to Florida.

Glacial Lakes. — The majority of existing lakes are the direct result of glaciation. This is well shown by the fact that most of the lakes of North America are concentrated within the northern half of the continent, precisely within the area which has recently been glaciated. Some of these lakes occupy depressions scoured or plucked out of the bedrock; still others are ponded back of dumps of glacial débris. Most are small, but some large lake basins as well are of glacial origin. The southern end of Lake Michigan owes its outline to a great crescentic ridge of morainal material behind which it was dammed; and many of the large lakes of the Alps are held in by moraines at the mouths of deep glaciated mountain valleys.

Lakes on Flood Plains. — Flood plains of streams, aside from the stream channels themselves, are areas of conspicuously poor drainage.

Shallow lakes are numerous in cut-off and abandoned meanders (*oxbow lakes*), and in the depressions between natural levees and the outer edges of flood plains. Lake Maurepas above New Orleans is a large lake of the latter type.

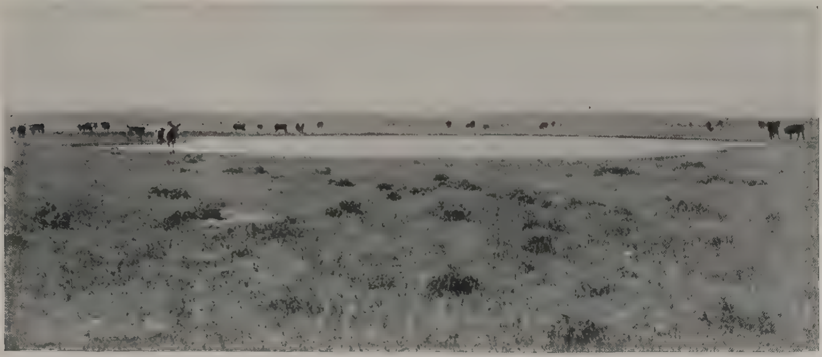


Fig. 109. — A temporary lake in a semiarid region. Wyoming. (U. S. Geol. Surv.)

Limestone Sinks. — Large areas in southwestern Kentucky and in southern Indiana are remarkable in that they are unglaciated uplands dotted with thousands of tiny lakes and ponds. These lakes occupy limestone sinks and are common also in parts of central Florida, Yucatan, Yugo-Slavia, and other limestone areas of the world. Sinks can contain water only (1) if their bottoms are below the water table or (2) if the outlets are clogged with insoluble clay left after solution of the limestone.

Lake Basins caused by the Wind. — In arid regions there are some intermittent lakes in shallow basins excavated by the wind. In more humid localities some of the hollows between live dunes also contain water which is prevented from sinking down through the sand by layers of decaying plant matter. Several large lakes of this kind are found among the dunes at the south end of Lake Michigan, and others are dotted about the Sahara.

Inherited Depressions. — The sea floor contains many irregularities and depressions, and therefore an area newly uplifted from beneath the sea often contains such depressions. Most of the Florida lakes which do not occupy sinks are of this sort. One of the largest is Lake Apopka near Kissimmee.

Playa Lakes. — The broad basins between desert mountain ranges are sometimes filled to a depth of a few feet after the sudden and violent rainfalls characteristic of such regions. Ephemeral or *playa lakes* are thus formed which disappear a few days or at most a few weeks after the rain has occurred (Fig. 110).

Relic Lakes. — Imbedded in the sands and clays of the basin of Lake Champlain, geologists find the shells of marine molluscs and the bones of seals and whales. This is doubly remarkable in that the fossil-bearing deposits now stand 440 feet above the sea. The same fossil remains, at even higher altitudes, are found also in the St. Lawrence valley as far west as Lake Ontario. From these facts it is evident that the St. Lawrence and Champlain valleys were occupied in comparatively recent times by an arm of the sea which has since been forced to withdraw because of uplift of the land. Such lakes as Champlain, relics of former seas, are called *relic lakes.* The Caspian Sea is the largest example of this type.

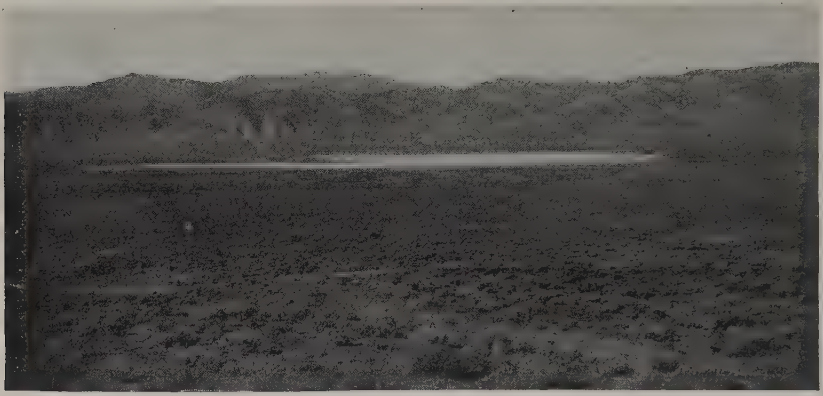


Fig. 110. — Playa lake in a desert valley near Ludwig, Nevada. The lake is rapidly disappearing through evaporation. (Knopf.)

LIFE HISTORY OF LAKES

Lakes are constantly coming into being through the agency of these basin-forming processes, and as constantly those in existence are being destroyed. The broad playa lake in a desert basin is the most ephemeral of all: the waters begin to be sucked up by evaporation as soon as they come to rest. But its containing basin is not destroyed; and with the next cloudburst, the lake reappears as before. In moist regions, however, evaporation goes on much more slowly, and at the same time the inflow of water into lakes is greater. Hence in these regions, lakes do not dry up. But here other forces are at work, which although they operate much more slowly, are far more dangerous to the existence of lakes because they destroy the basins which contain them. "Rivers are the mortal enemies of lakes" (Salisbury). The case of Lake Geneva and the river Rhone illustrates this striking statement. The lake of Geneva occupies a deep mountain valley and is at present about 40 miles long. It was originally 7 miles longer, but each year the turbid Rhone, which enters the lake from the east, brings down great quantities of fine sediment from the glaciers at its source and dumps this material in the quiet water, forming a great delta which fills the upper end of the lake basin from side to side. The water is nearly a thousand feet deep and the delta is correspondingly thick, but its bank-like front is creeping farther westward each year, and in time it will destroy the lake. While this is going on the water which spills from the surface of the lake at its lower end is cutting down its channel and is thus lowering the lake outlet. The water, as it passes under the many bridges of the city of Geneva, is beautifully blue and clear. All of the sediment from the upper Rhone has been added to the delta or has dropped to the bottom during the slow passage of currents down the lake. The lake has thus acted as a great settling-basin depriving the lower Rhone of tools with which to cut. Hence downcutting of the outlet is retarded, but it cannot be stopped. If these processes continue, the western rim of the lake basin will be lowered, and the water level will correspondingly drop, until the lake will be destroyed by this process if it has not already been filled up by the encroaching delta.

Downcutting of the rim of a large lake basin is more forcibly illustrated by Niagara Falls at the outlet of Lake Erie. The water at the brink of the falls is lowering its channel in a bed of limestone. At the same time the falls are retreating upstream at a rate which averages about 5 feet a year. When the falls have migrated up the river to a point opposite Buffalo, Lake Erie will have been largely drained. A lake not infrequently acquires a wholly new outlet through being tapped by a stream. An example seen by thousands of people every year is found in Yellowstone Lake near the center of Yellowstone Park. In former times this lake was larger than it is now and was drained at a point on its southwest rim by a stream leading through Snake River to the Pacific. The present Yellowstone River, flowing north, was then a small stream engaged in rapidly lengthening its valley headward on a steep slope. The head of this young valley eventually reached the lake, tapped and partially drained it, and diverted its waters into their present course through the Missouri to the Atlantic.

In the South Park of Colorado, in the district west of Pikes Peak, a stream is flowing through a valley whose sides are made of layers of tightly packed volcanic ash. Examination shows that these layers were deposited in a narrow lake, and that they came from an active volcano close by. The delicate ash, sifting down through the lake waters, carefully protected and perfectly preserved the remains of plants and animals that lived in and around the lake. The lake basin was gradually filled with ash and lava and was converted into a stream valley; and from the exposed layers of ash there have been taken more than 1000 species of insects, 250 species of plants (including numerous trees), fishes, and birds, all representative of the life of the time during which the lake existed. The filled-up lake, known as Lake Florissant, has thus become a storehouse of great scientific value.

It is thus apparent that all lakes will sooner or later be destroyed by (1) sedimentation in their basins or (2) drainage at their outlets, or both. To the former factor must be added the contributions made by wind-blown material, the accumulated bones and shells of lake-dwelling animals, and more important still, the remains of aquatic plants.

Climatic Control of Lakes

Lakes in Humid Regions. — A depression is formed in a certain locality. Will it contain a lake and if so what kind of a lake? In general the answer depends on rainfall and hence ultimately on climate. Each of the lakes with which we are familiar in eastern North America and in western Europe has streams or springs flowing into it, and an outlet or spillway determined by the lowest point in the rim of the containing basin. Since these regions are humid the rate of evaporation is relatively slow, and the constant inflow and outflow not only prevent the lake surface from fluctuating greatly but also keep the water fresh. In such lakes plant life is usually abundant, and the accumulations of decaying vegetable matter greatly help to fill up the lake basins, which are likely gradually to become ponds and swamps before they dry up completely.

Lakes in Arid Regions. — In the arid and semiarid regions of the world, where rainfall is slight and evaporation great, lakes rise and fall seasonally, and many dry up and disappear for months at a time. In fact the playa basins in the deserts which lie between the scattered ranges of Nevada, Utah, Arizona, New Mexico, and Sonora contain water for the most part only after sudden and infrequent rains. Here as well as in the desert of Gobi in central Asia, the great deserts of western Australia, the intermontane basins of the Andes, and many other arid regions, evaporation is so rapid and continuous, the soil so porous, and the ground water table so far below the surface, that the streams which start seaward, fed by the rains and snows of high mountains, disappear long before they reach their destination. Some of them dwindle away, spreading out their sediments in great barren flats covered with incrustations of salt and alkali which in wet seasons become shallow lakes and salt marshes. Other streams, where the water supply is greater, end in lakes. Wherever evaporation prevents the water from overflowing the rims of these interior basins, the lakes either are salt or are fast turning salt (Fig. 111). Every river carries dissolved salt in quantities too small to be detected by taste. Slowly but steadily this salt is added to the total already in the water of the lake. It cannot escape by overflow and it cannot escape by evaporation. So year by year the proportion of salt to water increases until not only can it be tasted but the water becomes undrinkable. The Great Salt Lake in Utah has a salinity of 18 per cent compared with less than $3\frac{1}{2}$ per cent for the ocean; its waters are so dense that bathers at the beach near Salt Lake City cannot sink, but float buoyantly upon the surface. As a swimmer wades back to the beach the water evaporates from his body, leaving it encrusted with tiny crystals of glistening salt. In the same manner slow shrinkage of the lake is leaving great dry beds of white salt to mark its former gently sloping shores. The chief salts are

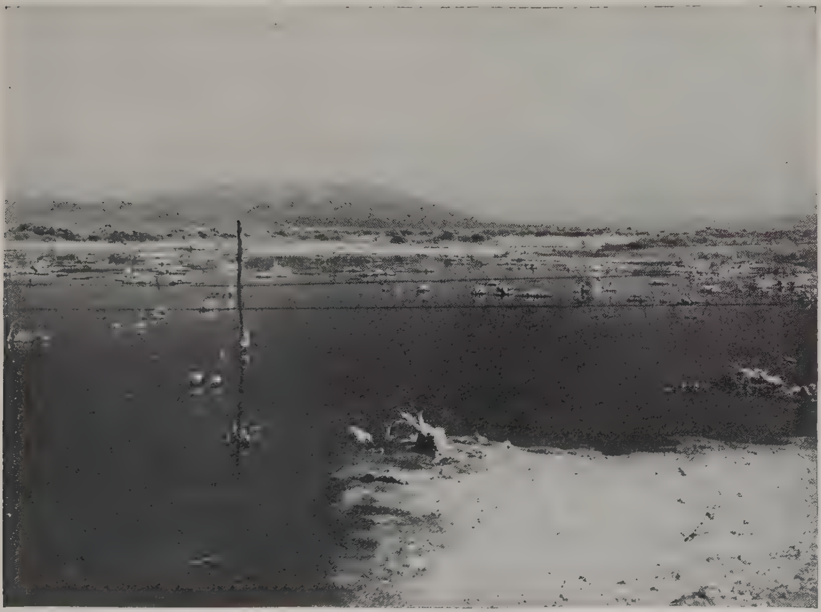


Fig. 111. — Alkaline salt lake near Parma, Colo. (U. S. Geol. Surv.)

common salt (sodium chloride) and sodium sulphate. Calcium carbonate is also present and is commonly deposited as granules on the bottom and shores.

Extinct Lakes. — Observant travelers in Utah and Nevada have had their curiosity aroused by frequent sights of parallel rows of cliffs and great flat-topped terraces forming huge flights of steps up the mountain sides to levels as high as 1000 feet above the valley floors (Fig. 113). When examined in detail, these " steps " prove to be wave-cut cliffs and wave-built terraces, bars, and beaches, as well as deltas high and dry. These features are arranged in sets at different levels,

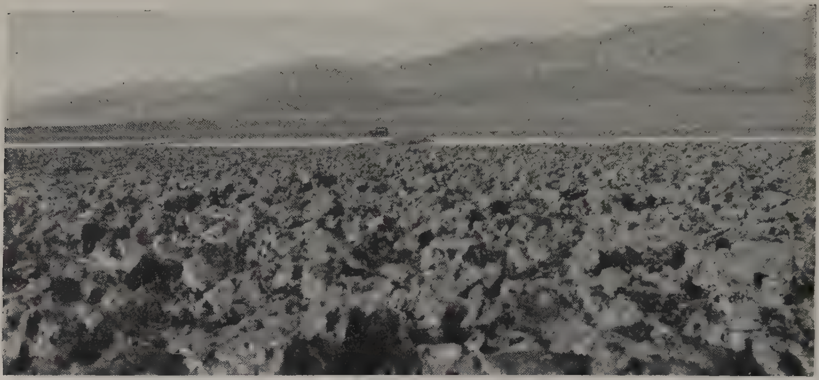


Fig. 112. — Salt deposits on the floor of Death Valley, Calif. The salt in the foreground is etched into pinnacles by weathering processes, and is covered with wind-blown dust. Salt formed recently is white. The valley floor is here 250 feet below sea level. The car in the middle distance gives the scale. (Longwell.)

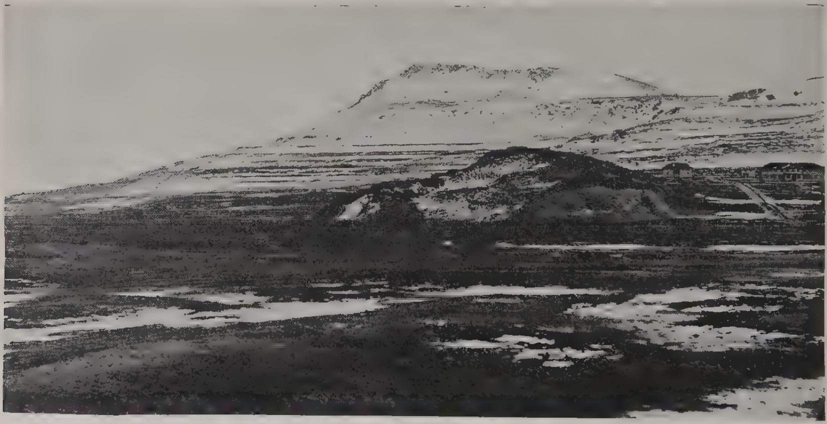


Fig. 113. — Former shore lines and wave-cut terraces of the ancient Lake Bonneville.

and when traced along the various mountain ranges of the region, the highest set is found to mark the shoreline of a lake which had an area of 20,000 square miles (nearly as great as that of Lake Michigan) and a maximum depth of more than 1000 feet. The water must have been 850 feet deep at one time above the site of the Mormon Temple in Salt Lake City. The outlet was to the north, through Red Rock Pass to the Snake River, and thence through the Columbia to the Pacific. Since there are several sets of shorelines at different levels, we must conclude that the lake dwindled by stages from its former gigantic size to that of the Great Salt Lake, which represents the deepest pool in the bottom of the basin of its predecessor (Fig. 114). Similarly, Pyramid Lake and its neighbors in Nevada are the residual pools of another huge water body, which rivaled in size the present Lake Erie. The Utah lake has been named *Bonneville* and the Nevada lake *Lahontan.* The fact that some of the terraces of Lake Bonneville are associated with glacial deposits indicates that the lake had its origin in the Glacial Period. Salt deposits interbedded with the lake clays, all now high and dry, bear witness to the evaporation which has been going on during the time since the Glacial Period.

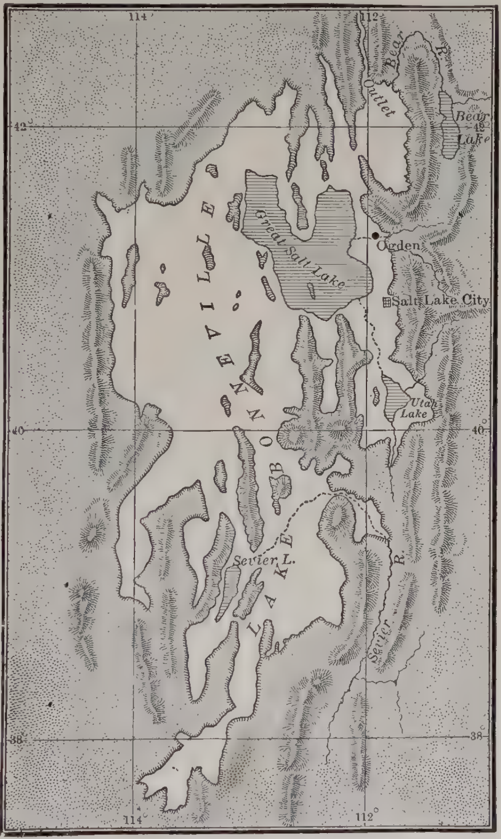


Fig. 114. — Map of the former Lake Bonneville; ruled areas show present water-bodies.

The Salton Sea. — The main line of the Southern Pacific Railroad eastward from Los Angeles has the curious distinction of running for more than 60 miles through an area that lies below the level of the sea. The area is the Salton Basin. It is a basin without outlet, and its bottom is 273 feet below sea level. In its center is a shallow salt lake known as the Salton Sea (Fig. 115). In the recent geologic past this area sank below sea level and would have been filled with sea water; but Colorado River, heavy with silt, built up a combined delta and fan so rapidly that it maintained a dam across the head of the Gulf of California. Therefore the subsiding Salton Basin remained dry, except for fresh water that sometimes ran into it when Colorado River shifted its course on the growing fan. Like other deltas and alluvial fans, this one was built up by shifting distributary channels of the river, which sometimes discharged water into the head of the Gulf and occasionally into the isolated basin. When dry, the basin was kept virtually a desert under the prevailing arid climate. The last natural discharge into the basin occurred perhaps 300 to 1000 years ago. In 1900 the trapped water had dwindled to a small salt lake. At this time the fertility of the alluvial soil was realized, the basin area was settled, and an irrigation canal was dug from Colorado River. Since the floor of the basin was well below the baselevel of the river, the gradient of the canal was greater than the gradient of the river between its mouth and the head of the canal. The flow of water into the canal was inadequately

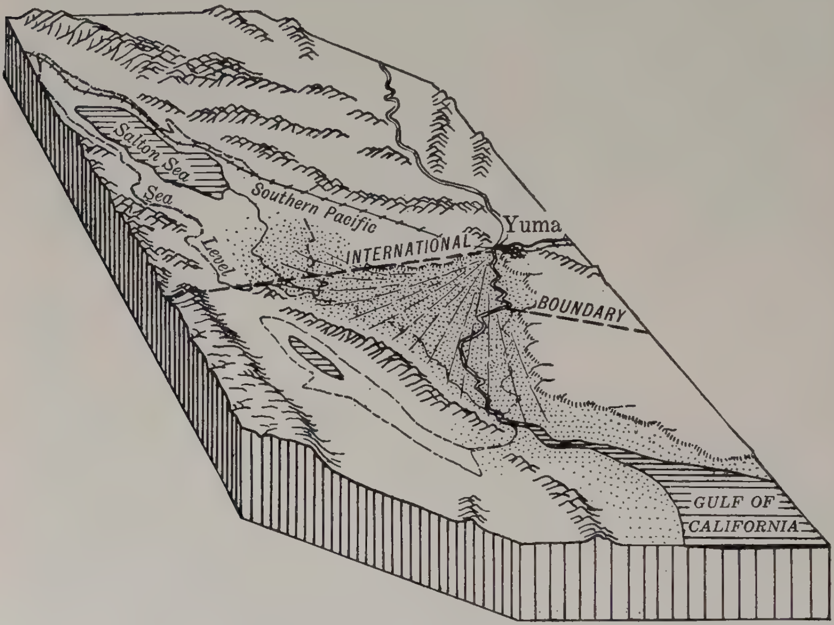


Fig. 115. — Delta of the Colorado River and the Salton Basin. Dashed lines show sea level. Longest dimension of block about 200 miles.

controlled by headworks designed to be protective. In 1905 the Colorado rose in flood, overtopped the headworks, poured into the canal, and following the new steep gradient, cut a great trench, in some places 80 feet deep, swelled the Salton Sea to many times its former size and depth, and flooded the railroad right of way for more than 40 miles. The inundation of the basin resisted all efforts at permanent control for more than 18 months; but it was finally mastered, and the railway tracks were shifted as a precaution against possible later floods.

Indirect Functions of Lakes

Large lakes are effective in modifying local climates by increasing atmospheric humidity and by cooling the air in summer and warming it in winter. Lakes of all sizes are very important as regulators of stream flow, acting as storage reservoirs and minimizing the height of floods in lower regions to which they are tributary. This fact has been recognized by the Egyptian Government in its project to increase artificially the size of Lake Tana in Abyssinia, one of the important sources of the Nile, in order to increase its effectiveness as a regulator of water supply in the Lower Nile Valley, an area of great economic importance. Similarly, suggestions have been made in the United States that a series of artificial lakes be constructed in the Mississippi drainage basin in the attempt to control near their sources such floods as caused the disaster of 1927. All lakes likewise act as settling basins for river sediment. Most of the streams tributary to Lake Erie are well loaded with mud, but the clarity of the water which pours out of the lake and over Niagara Falls bears testimony to the amount of material which is constantly being dropped upon the lake bottom.

SWAMPS

Swamps are areas of saturated ground. The majority of swamps represent a stage intermediate between lakes or ponds and dry land. Most lakes will in time become swamps, and many shallow basins alternately contain swamps and lakes according to the season. Swamps are commonly found in three types of regions, but these regions by no means exhaust the possibilities. (1) Swamps are both numerous and large on nearly level coastal plains which are former sea floors slightly uplifted. Such swamps are distinguished from salt marshes and are almost continuous along the South Atlantic and Gulf coasts of the United States, chief among them being the Dismal Swamp in Virginia and North Carolina, and the Everglades in southern Florida. Some of these swamps may be old lagoons, uplifted together with the bars by which they were shut off from the sea. (2) Flood plains and deltas with their basins formed by old channels and by natural levees contain great areas of swamp land. (3) Broad glaciated areas such as the greater part of the Great Lakes region of the United States, northeastern Canada, and the Baltic Plain of northern Germany are dotted with swamps, most of them small.

Formation of Peat. — In humid regions, the shores of small lakes and protected coves of large lakes support an abundance of aquatic vegetation such as pond lilies, water weeds, and rushes. As these plants die, their substance decays in the water by slow oxidation, which is caused largely by the activities of bacteria. During the metabolism of the bacteria, waste products, antiseptic in their action, are excreted; hence when this waste matter reaches a certain concentration in the lake water the bacteria can no longer exist, further decay is prevented, and the partially decomposed matter is preserved. This matter is brownish or blackish, has a high carbon content, and is known as *peat.*

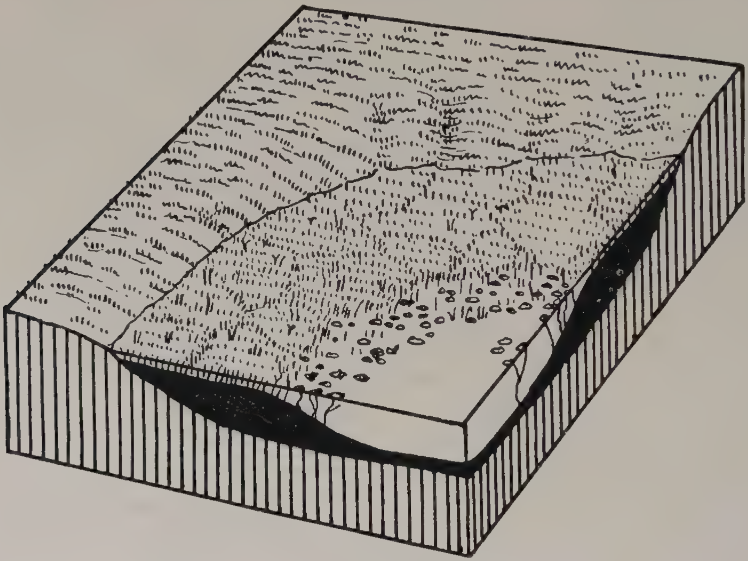


Fig. 116. — Destruction of a small lake by formation of peat. The accumulating layer of peat (solid black) is fringed by aquatic vegetation consisting of water weeds and pond lilies while directly above it are encroaching semi-aquatic plants, mosses, and bushes.

As peat is gradually built up along the shore of a lake, newer generations of aquatic plants advance toward the center of the lake, and other types of vegetation such as mosses encroach over the peaty area which was formerly water. Thus the lake, surrounded by concentric belts of different kinds of plants, gradually decreases in size until it is obliterated, and a bog floored with a thick accumulation of peat takes its place (Fig. 116).

Economic Aspects of Peat. — In many countries, especially in Europe, peat is cut from the bogs, dried, and used as a domestic fuel. In America peat has hitherto been little used because of an abundance of coal and wood. The peat resources in bog lands within the United States, however, are enormous, and will probably constitute a valuable source of fuel in the future.

Peat is of special interest to the geologist in that it represents the first stage in the transformation of vegetable matter into coal. All gradations may be observed from peat through lignite, bituminous (" soft ") coal, and anthracite (" hard ") coal. Thus if present peat bogs, the best of which lie in temperate and cold humid regions, were left untouched, many of them in the course of time would be covered with sediment, the remaining necessary changes would take place, and the result would be the formation of coal interbedded with the rocks of the Earth's crust.

Economic Value of Swamp Lands. — Peat is not the only valuable deposit of swamp origin. In lakes and swamps as well as in warm springs and even in the surface waters of the ocean live great swarms of microscopic plants called *diatoms,* which secrete minute shells of silica. When these organisms die, their tiny shells fall to the bottom and build up a white, porous deposit known as *diatomaceous earth.* Beds of this material, some of them hundreds of feet in thickness, are now excavated and used in the manufacture of dynamite, polishing powder, and other products.

Certain valuable deposits of iron ore can be traced back to their origin in lakes and swamps. Many areas of marsh and standing water today are inhabited by countless numbers of microscopic organisms known as *iron bacteria,* living wherever the water contains dissolved iron. As an essential part of their life process, they secrete the iron from solution, convert it into insoluble form, and precipitate it, causing an accumulation of iron (usually in the form of limonite) on the lake or swamp floor. The bacteria, though minute, exist in such inconceivable numbers, and in successive generations work through such long periods of time, that large deposits of iron are gradually built up. Such deposits of " bog iron ore " are particularly abundant in the glaciated regions of Europe, Asia, and North America, although they are now of very slight importance in the mining industry.

The agricultural value of swamp land when drained is not to be

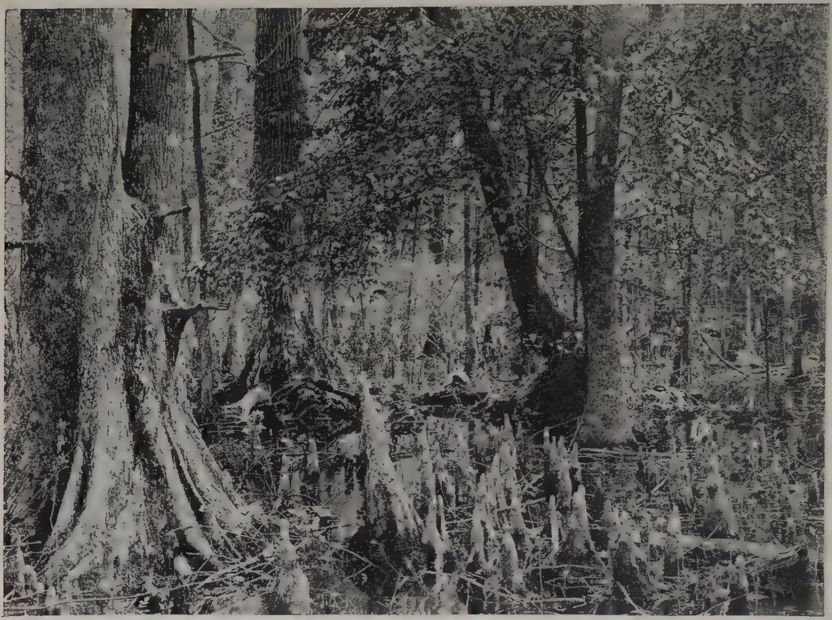


Fig. 117. — Dismal Swamp, Va. The projections from the cypress roots serve to give them air; they extend downward into the mud and help to anchor the tree in the semi-liquid mass of the bog. (U. S. Geol. Surv.)

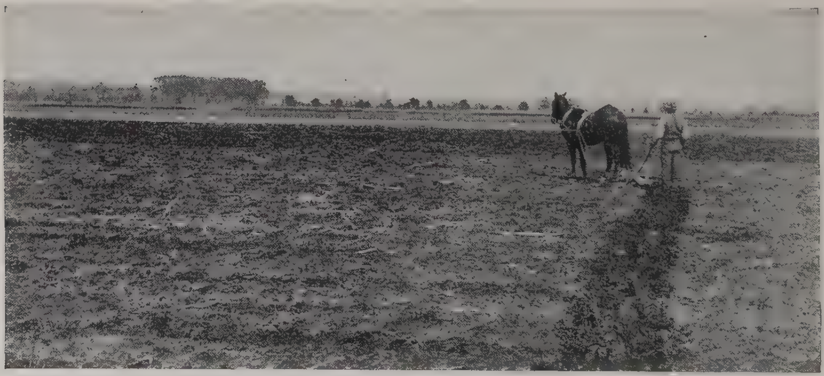


Fig. 118. — Reclaimed land from the Dismal Swamp, Va. (U. S. Geol. Surv.)

minimized since swamp soils are highly fertile. The swamps and bog lands in the United States have a combined area greater than the area of New England, and with proper draining much of this land could be reclaimed for farming. Drainage canals and ditches have been used for centuries in Flanders and Holland; and now in many parts of the United States large areas of wet land are being prepared in this way for agricultural use (Fig. 118). Thus artificial means are being used to accomplish what in time would have been achieved by normal erosion — more complete gradation of the land.

CHAPTER VIII

OCEANS AND SEAS

General Features

Geologic Rôle of the Oceans and Seas. — Marine waters cover about three-quarters of the globe. Their influence on the geologic history of the Earth's surface has been both direct and indirect. The importance of the oceans in regulating climate, for example, is well known. It is a remarkable fact that throughout the hundreds of millions of years of geologic history, the general temperature of the Earth has never fallen below the freezing point nor exceeded the boiling point of water. For this remarkable constancy we must thank the unflagging energy of the sun's warmth, but the oceans have also undoubtedly served as a great stabilizer by storing up excess warmth against times of lessened solar radiation. Ocean currents carrying warm waters into the higher latitudes and returning the colder waters toward the equator also aid greatly in distributing the heat and softening the contrasts between climatic zones. Moreover, evaporation from the ocean's surface in the last analysis supplies all of the moisture borne by the winds to fall as rain and snow upon the lands. The intense aridity of the basins of the several continents, as the great desert of Central Asia, the deserts of Australia, the Sahara, and the Great Basin of the western United States, suggest what might be expected if the Earth had limited oceans.

Through its waves and currents the marine water is an energetic agent of erosion, gnawing away relentlessly at the margins of all the lands. Moreover, the seas and the oceans are the final settling reservoir in which are deposited the sediments brought down by the rivers, as well as those produced by marine erosion. Upon the sea floor these sediments are shifted about to fill the depressions, or, in the shallow places, tossed up to build out the coast lines of the lands and ultimately to be compacted and cemented into sedimentary rocks.

Relations of Oceans and Continents. — If the waters were withdrawn from the ocean basins, we should see that the grandest relief features of Earth's rocky crust are not its mountain ranges but the uplifted continental masses that lie as vast plateaus 3 miles above the enormous plains of the ocean floor. The naked face of the moon presents to us a spectacle of this sort, for its maria or "seas" would be such in reality if there were water upon its surface.

Presumably the continental masses stand high because they are made of lighter, granitic rocks and the oceanic areas are depressed because they are formed of heavier, basaltic rocks. On the moon, the depressed segments are relatively small; but Earth's oceans are far greater than its continental areas. It is apparently only a coincidence that there is just sufficient water on the Earth to fill the ocean basins brim full and place the shoreline upon the margin of the continents.

Size of the Oceans. — Marine waters attain a maximum depth of more than 6 miles (35,410 feet) and an average of about 2½ miles (13,000 feet), whereas the average elevation of the emergent continents is only about half a mile above sea level. So vast is the volume of oceanic waters that if the continents were planed down and leveled into the basins, a universal ocean would cover the entire Earth to a depth of more than 1½ miles. It is important to realize, however, that even this depth is slight in comparison with the diameter of the Earth. If, for example, a globe 3 feet in diameter were dipped into water and then withdrawn, the film of wetness adhering to it would represent to true scale an ocean half a mile in depth, and if, in drying, the globe should warp by so small an amount as to lessen its diameter at any place by one-hundredth of an inch, the change would correspond to the depression of one of the major ocean basins. It is evident, therefore, that relatively trivial warping of the Earth's crust as a whole would suffice to deepen parts of the ocean basins and draw off the water from the continents or, by elevating parts of the ocean floor, to cause an overflow of the lower land areas, with vast marine inundations. Whatever the cause, such changes have occurred many times in the geologic past.

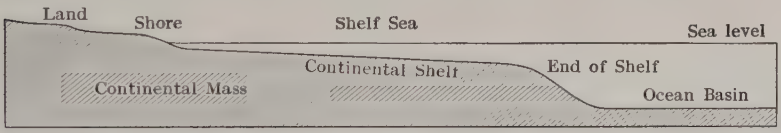


Fig. 119. — Section through edge of continent into ocean basin. Vertical scale greatly exaggerated, causing slopes to appear much steeper than they actually are.

At present the oceans more than fill their basins and flood the margins of all the continents. These slightly submerged borders of the continental masses are known as continental shelves (Fig. 119). They are widened on the one hand through landward planation by the sea and on the other through the deposition of sediments swept toward the ocean by marine currents. They are broad along stable coasts where these processes have operated without interruption for long geologic ages, but commonly are narrow where young mountains have been formed near the continental margin. Thus, for example, the continental shelf is 60 to 80 miles wide on the Atlantic coast off the Carolinas and only 10 to 25 miles, or even less, along the Pacific coast of California.

Oceans Versus Seas. — In common usage the words sea and ocean are synonymous, and we speak of deep-sea deposits, and of going to sea, when actually we have in mind the ocean. There is, however, a technical distinction observed by scientists who apply the term *ocean* only to those four vast bodies of deep water that occupy the ocean basins proper and lie *between* the continental masses. The seas, on the contrary, are relatively shallow bodies of marine water that lie *upon* the continental platforms as, for example, the North Sea or Hudson Bay. This technical restriction of the term sea accords with its original use by the peoples of northwestern Europe, who applied it to the North Sea and the Baltic in contradistinction to the Atlantic or outer ocean.

The distinction is one of great significance because of the profound influence of depth on the processes that operate on the ocean floors. The seas are for the most part less than 600 feet deep, whereas the average depth of the oceans is more than twenty times as great. The floor of the seas is a region of activity, touched by the warmth of sunlight and stirred by waves and currents that shift its sediments and bring oxygen and land-derived food to its teeming life. It has been the cradle of evolution for the lower forms of life and the scene of geologic processes of profound importance. The great oceanic floor, on the contrary, is a realm of dreary monotony. Except where submarine volcanoes or earthquake lines are active, nothing disturbs its quiet save the noiseless descent of wind-blown or meteoric dust or of derelicts of the surface life miles above; and its denizens eke out an existence in utter darkness and cold.

The Mediterranean " Sea " fits the definition of neither sea nor ocean, for it is of oceanic depth and yet is comparable in size to a sea and closely circumscribed by lands. The term *mediterranean* is therefore used in a generic sense to include such circumscribed bodies of deep water lying between continents. The Caribbean " Sea " is thus another mediterranean.

Classes of Seas. — The seas may be arranged, for convenience of reference, in three classes: marginal, epeiric, and relic.

Marginal seas are those which lie upon the continental shelf and are more or less openly connected with the ocean. Where they are somewhat delimited by projecting lands, various portions of the marginal seas bear distinct names; as the North Sea, the Yellow Sea, the Gulf of St. Lawrence, Cape Cod Bay, or the Gulf of Maine. The shallow water overlying the continental shelf of the central and southern Atlantic coast of the United States is just as much a sea, but it and others similar have not been named.

The *epeiric seas* [Gr. *epeiros*, a continent] are those that lie so far in upon the continent as to be largely land-locked. At the present time there are but two good examples, the Baltic Sea and Hudson Bay, but during certain past geologic ages, when the lower portions of the continents became widely flooded, epeiric seas were of vast extent and great importance. In fact most of the sedimentary rocks upon the present lands were formed in epeiric seas.

The *relic seas* are those like the Caspian, which, through crustal uplift, have become isolated from the oceans. They are, indeed, great brackish lakes and should be so called; but their former marine connections are indicated by the life they still harbor. For example, although the waters of the Caspian are now almost fresh, due to the inflow of great rivers, its fauna includes marine fishes, porpoises, and seals. It is known to have been connected with the Arctic Ocean as well as the Mediterranean within rather recent geologic time.

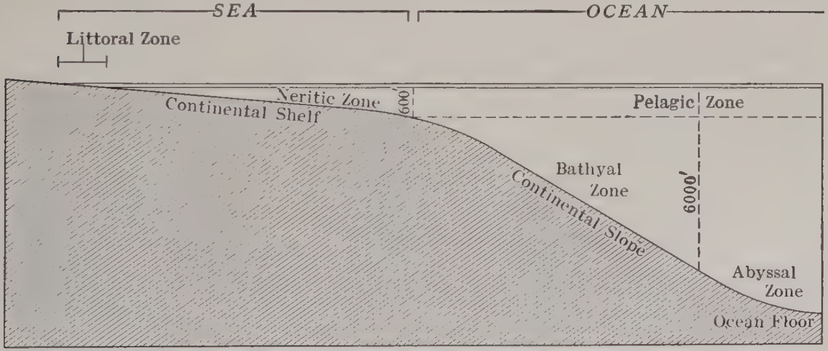


Fig. 120. — To show the several depth zones of the marine realm. The vertical scale and the steepness of slopes are much exaggerated, and the full depth of the ocean is not shown. High and low tide levels are indicated by the two lines at the top.

Depth Zones. — The depth of the water is one of the most important factors in the marine environment for it conditions the amount of light and warmth so vital to marine life, and also the effectiveness of the waves and currents that determine the character of the sediments. From this point of view, the marine realm may be subdivided into several bathymetric or depth zones (Fig. 120).

The *littoral zone* extends between extremes of high and low tides and includes, besides the actual shore slope, the mud flats, laid bare at low tide, and the salt-water marshes, which are flooded only at highest tide. The entire width of the littoral zone is usually not over 2 miles; but in a few regions of exceptionally high tides or of rapid silting, it reaches a considerable width. For example in the Bay of Fundy, where the average tide rises between 30 and 40 feet, the mud flats at ebb tide are 4 or 5 miles wide. In parts of the Mississippi River delta the salt-water marshes attain a width of fully 25 miles.

The environment of the littoral zone is more varied than that of any other part of the sea floor. Alternately covered by salt water and laid bare to the atmosphere, it may be baked in sunshine one hour, and deluged by the fresh water of a storm the next. Moreover, the waves break here with their greatest vigor, and wave and current action is strongest. At any one place the environment changes with the tides and storms from day to day, and it is equally variable from place to place along the shoreline, depending on the degree of exposure to the open sea. Accordingly the littoral region is the most difficult life zone, and relatively few kinds of animals live here though some, like the barnacles and crabs, exist in vast numbers.

The *neritic zone* includes the shallow, sublittoral sea floor from the limit of low tide out to the margin of the continental shelf; that is, to a depth of about 600 feet. It is known to the Germans as the Flachsee (flat sea) and, indeed, its floor is in general remarkably even, for it is mantled by the sediments derived from the land and shifted by the waves and currents into the deeper places, smoothing out the inequalities of its surface. The average width of the neritic zone being about 75 miles, its seaward slope averages less than 10 feet to the mile, and it would appear to the eye as an utterly monotonous level plain. Only locally, and generally near the shore, does it reach an inclination as great as 50 feet to the mile.

The neritic zone is a realm of change and activity. Waves and currents keep the water in motion, salinity and muddiness vary from place to place, and the temperature changes with the seasons and varies with the latitude. Sunlight penetrates to the bottom. It is the most hospitable and stimulating environment for sea life and the teeming host of creatures that live here play an important geologic rôle.

Beyond the edge of the continental shelf there is a somewhat steeper incline, known as the *continental slope*, that is fairly well defined from a depth of 600 to about 6000 feet, where it imperceptibly grades into the abyss of the deep ocean. The intermediate depths that cover the continental slope constitute the *bathyal zone*. Due to the greater depth, bottom currents are feeble here and no coarse sediment can be introduced; but fine, land-derived muds cover a vast expanse.

The *abyssal zone* includes the whole of the ocean bottom below a depth of 6000 feet. Its average depth is about 13,000 feet and in general it is monotonously flat, lacking the smaller relief features such as the hills and valleys of the land. Nevertheless on a large scale there are swells and depressions rising above and sinking below the general level of the ocean floor. The depressed areas, if more than 18,000 feet below sea level, are known as *deeps* and of these 57 have now been discovered.

The floor of the deep ocean presents an inhospitable environment for living things, since at this depth no sunlight ever penetrates the utter darkness; the pressure amounts to about 1 ton per square inch for each mile of depth; and the temperature is less than 4° C. Life is sparse and grotesquely specialized. In the absence of sunlight, no plant life exists and the animals are either carnivores or else scavengers, feeding on the dead organisms that settle down from the surface layers of the ocean. Over the abyssal floor, land-derived sediments are wanting, but everywhere there is a covering of fine, soft deposits known as *oozes*.

The surface waters of the open oceans constitute the *pelagic zone* or realm [Gr. *pelagos*, the open sea]. The organisms that live here enjoy the warmth and sunlight but must perpetually swim or float. As indicated above, they may contribute to the bottom deposits of the ocean basins.

Composition of Marine Water. — A barrel of sea water contains about 12 pounds of dissolved mineral salts. These dissolved salts constitute about 3½ per cent of the weight of marine waters. Chemical analyses from all parts of the oceans, and from different depths, show that the composition of the water in the open ocean is remarkably uniform, and that over 99 per cent of the dissolved mineral matter probably represents only a half dozen common salts, as follows:

Sodium chloride, $NaCl$	77.8 per cent
Magnesium chloride, $MgCl_2$	10.9 per cent
Magnesium sulphate, $MgSO_4$	4.7 per cent
Calcium sulphate, $CaSO_4$	3.6 per cent
Potassium sulphate, K_2SO_4	2.5 per cent
Calcium carbonate, $CaCO_3$	0.3 per cent
Minor constituents	0.2 per cent
	100.0 per cent

Although common salt ($NaCl$) constitutes more than three-quarters of the dissolved mineral matter, there is little more than a trace of calcium carbonate ($CaCO_3$), the ratio of salt to calcium carbonate being almost 260 to 1. This relation is the more striking when it is recalled that river water, whence the oceanic salts have been derived, normally carries far more calcium carbonate than sodium chloride. The discrepancy is due, of course, to the fact that $CaCO_3$ is either used by organisms or chemically precipitated about as fast as it is delivered to the sea, whereas nearly all the salt (NaCl) that has been brought down to the sea since the beginning of time remains in solution. The total quantity of the dissolved salts is astonishing, for it amounts to about 32,000 million million tons, and if precipitated and crystallized into a bed of solid salt, would be sufficient to cover the whole of the United States to a depth of more than $1\frac{1}{2}$ miles.

Silica and nitrogen, though present only in traces in marine waters, are of vast biologic consequence. The nitrogen is required as food and the silica as shell material by the single-celled plants known as diatoms, which float near the surface in incredible numbers, and form, to a large degree, the ultimate food supply for all the animals of the marine realm. Diatoms flourish and multiply rapidly if the supply of dissolved nitrogen and silica is enriched but their expansion is checked by a falling off of these foods, so that the " pastures of the sea " may be said to depend on the supply of nitrogen and silica. The latter is one of the common minerals in river water and its rarity in the sea is probably the result of its use and precipitation by organisms.

In addition to the mineral salts, marine water holds in solution oxygen, nitrogen, carbon dioxide (CO_2), and other gases. As a rough average, 1000 cu. cm. of sea water holds in solution about 20 cu. cm. of air (oxygen and nitrogen) and 23 cu. cm. of carbon dioxide, both measured at standard temperature and pressure. The importance of these gases is out of all proportion to their quantity. All the animal life in the seas and oceans requires oxygen. Oxygen is necessary, moreover, for the oxidation and destruction of the carcasses of dead organisms on the ocean floor, and it thus aids in keeping the water clean and habitable. In places, such as the depths of the Black Sea, where defective vertical circulation fails to provide a sufficient supply of oxygen, the water becomes putrid and foul with the gases of decay, and only anaërobic bacteria are able to live. The entire ocean bottom would be equally forbidding if it were not for the bountiful supply of dissolved oxygen.

Carbon dioxide is the primary food stuff from which green plants draw their sustenance and in turn build up the organic foods upon which animal life is dependent. Without it no life would be possible. At the same time it aids in the solution of the calcium salt, $CaCO_3$, as explained in Chapter III. Since the solubility of CO_2 varies inversely as the temperature, the cold waters of the polar regions and of the deeper ocean bottom can hold more CO_2 than the warmer surface waters. It is this excess of CO_2 in the abyssal depths that causes the solution of all limy shells that settle there.

Movements of Marine Waters

Waves. — The common movements of marine water include the waves, tides, and a variety of currents. Probably the most important and the most direct of these in their geologic effect are the waves.

The ordinary waves of the sea are generated by the wind blowing in irregular gusts and pressing unevenly upon the surface of the water, which is thereby thrown into little undulations. Once formed, these undulations are maintained and increased by the pressure against their windward side and so are driven forward in endless succession.

It is important to realize that the *wave form* travels ahead, independent of the movement of the water itself, just as waves may ripple across a field of standing grain though the individual stalks merely bow as each wave passes and then return to their original positions. Indeed, if the water actually rushed forward with the velocity of storm waves, the oceans would hardly be navigable. The path of movement of an individual particle of water is almost a closed circle, for it rises and rides forward with the crest of the wave only to slide back into the next trough to its original position, a fact that can be observed by watching a bit of cork or driftwood as the waves pass under it. As a matter of fact, due to the friction of the wind, it may not return quite to its original position but may be slightly advanced by each wave. In this manner, *wave-formed currents* are generated.

The height (i.e., vertical elevation of crests above trough) and the length (i.e., horizontal distance from crest to crest) of the waves increase with the velocity of the wind, its duration in a given direction, and the length of " fetch " across open water. For this reason small bodies of water and protected embayments of the coast are never affected by great waves. A distance of 400 or 500 miles is sufficient to produce the greatest waves, however, since the wind does not blow steadily in any one direction over greater distances.

Size and Velocity of Storm Waves. — Observations have shown that storm waves in the North Atlantic commonly have a length of 400 feet and a height of 20 feet, but in times of exceptional storms they may attain a length of over 1000 feet and a height of more than 40 feet. Captain Tanner of the United States Navy has photographed a storm wave exceeding 50 feet in height. The velocity of the waves increases with their length and is about 25 knots, for example, when the waves are 400 feet long and 40 knots for those 1000 feet long.

When great waves run out of the storm they decrease in height and become more rounded, but they may continue for hundreds or thousands of miles as the long heavy undulations of the surface known as *"ground swells."* In this form the waves may attain a length exceeding 2000 feet.

The wave motion decreases rapidly in depth and, according to theory, the orbit of movement of a particle at a depth of one wave length should be only 1/534 of that at the surface. That is to say, in storm waves 20 feet high and 600 feet long, the surface particles of water will move through orbits having a diameter of 20 feet and those at a depth of 600 feet through an orbit only $\frac{2}{5}$ of an inch in diameter. The sea floors off exposed coasts are generally affected by waves to a depth of 200 or 300 feet, and exceptional waves move fine sediments to the edge of the continental shelf, a depth of about 600 feet.

When a wave passes into shallow water, an important change takes place in its form and behavior. The wave becomes higher and shorter and its front side steeper and more deeply concave until the crest arches forward, unsupported, and collapses with a roar. These collapsing waves or *breakers* form the *surf*, so common a feature along coasts. The height of the wave determines the depth at which it will break, and thus small waves break in shallow water, near shore, whereas the great storm waves break where the water is from 10 to 20 feet deep.

Tides. — Due to the attraction of the moon, modified by that of the sun, the ocean surface is lifted into two vast but low tidal bulges, one on each side of the Earth. These bulges remain fixed with respect to a line extending to the moon, but since the Earth in its daily rotation turns under them from west to east they seem to run round the Earth from east to west.

In the open ocean, the surface merely rises a little and then subsides again as each tidal bulge passes; but where the bulge impinges against a coast the water is dragged forward, piling up on the shore and then receding, thus producing the familiar phenomenon of the *tides*. The height of the tide is determined largely by the configuration of the coast. On open, exposed coasts it is not more than 6 or 8 feet, and in protected embayments as, for example, the Gulf of Mexico, it is only 1 or 2 feet; but in estuaries that open out toward the advancing tide, the water rises higher as it converges, bringing about exceptional conditions like those in the Bay of Fundy, where the tide rises normally 30 to 40 feet and exceptionally as much as 50 feet (Fig. 121).

The immense bodies of water thus moving in and out of bays and estuaries and along the coast every 6 hours, produce strong tidal currents, which, like rivers, have a two-fold geologic function in that they both erode and transport. The work done by the tides in scouring the bottom and transporting material along some coasts is very great. The rise and fall of the tides indirectly aids in the attack of the waves on the land by increasing the vertical range of their contact with the lands.

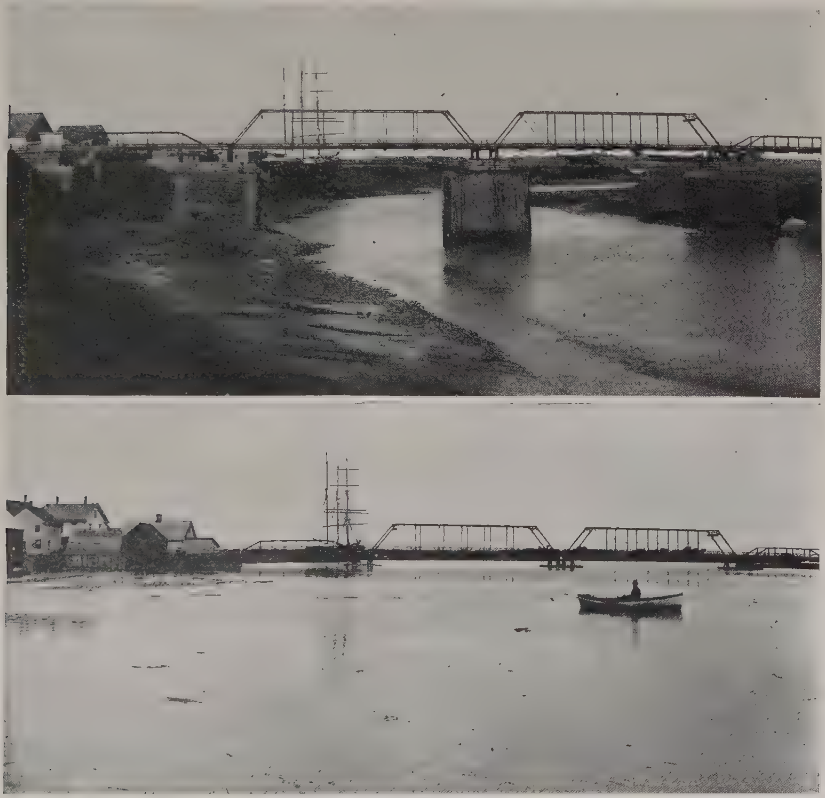


Fig. 121. — High and low tide in the Bay of Fundy. Note the same bridge and sailing vessel in both views. Port Williams, Nova Scotia.

Currents. — There are several types of marine currents, only the chief of which need be considered here. *Surface currents* are produced by the friction of the wind when it blows persistently in one direction for a time. Where they strike the coast and, with the aid of the breaking waves, tend to pile the water up on the shore, there is a compensating bottom current, the *undertow*, flowing seaward (Fig. 122). On a straight coast, with a gently shelving sea floor, the undertow is so diffused as to be very gentle, but where irregularities in the bottom constrict the returning current to more definite channels, the undertow may be a menace to swimmers and a powerful agent in sweeping sediments seaward into deeper water. Where the waves strike the shore obliquely, the run of the water due to successive impulses of the waves generates a current parallel to the shore, known as the *littoral* or *shore* current (Fig. 123). Such currents sweeping material along the coast are responsible for the formation of beaches, spits, and bars. The fine sand of Daytona Beach, Florida, for example, comes not from the interior of that limestone country, but along the shore from far to the north, where the rivers of Georgia and the Carolinas have brought it down to the sea out of the southern Appalachians.

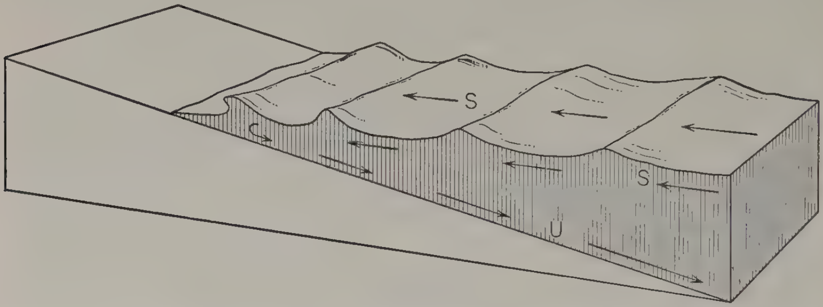


Fig. 122. — To show the relation of the wave-formed surface current (*S*) and the undertow (*U*) when the waves are coming directly onshore.

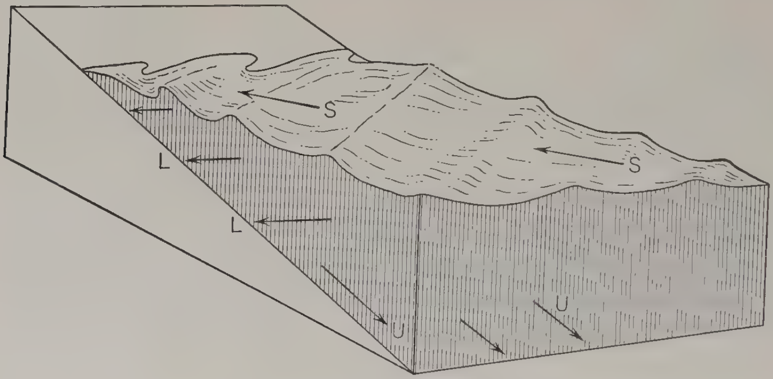


Fig. 123. — To show the relation of the wave-formed surface current (*S*) to the littoral current (*L*) and the undertow (*U*) when the waves are oblique to the shore.

Tidal currents are strong where the tide sweeps in and out of deep estuaries or bays. Where the shoreline is oblique to the east-west path of the tidal bulge, there may be a strong component of the tidal current running parallel to the shore. This is true, for example, along the western coast of Newfoundland, where the ebbing tide sets to the northeast through the Strait of Belle Isle with a velocity of 4 or 5 miles per hour, a rate that compares well with the flow of large rivers.

The more general *ocean currents* have for the most part only an indirect geologic effect by modifying the climate of the lands, for they rarely touch the shores or the bottom with sufficient velocity to erode or transport sediments. On the other hand, they have a distinct geographic value in drifting organic waifs to strange lands. These currents are generated by several different factors. The unequal heating of the ocean by the sun in tropical and polar regions would establish a slow general circulation of its waters through convective movements. This action, however, is controlled, hastened, and magnified by the wind belts of the Earth, and by the disposition of land and sea. Driven by the trade winds, there is, in the equatorial regions of the Atlantic as well as the Pacific, a broad current moving westward along the surface. When this strikes the continental coasts it divides, one part turning northward, the other southward; and each, circling, returns to the equatorial belt, thus making in each ocean a vast eddy, one north, the other south of the equator (Fig. 124). When these broad slow movements, which are known as *drifts*, approach the coasts, the water tends to accumulate, and where confined by the configuration of the land is hastened in its motion, giving rise to *streams*. Thus in the North At-

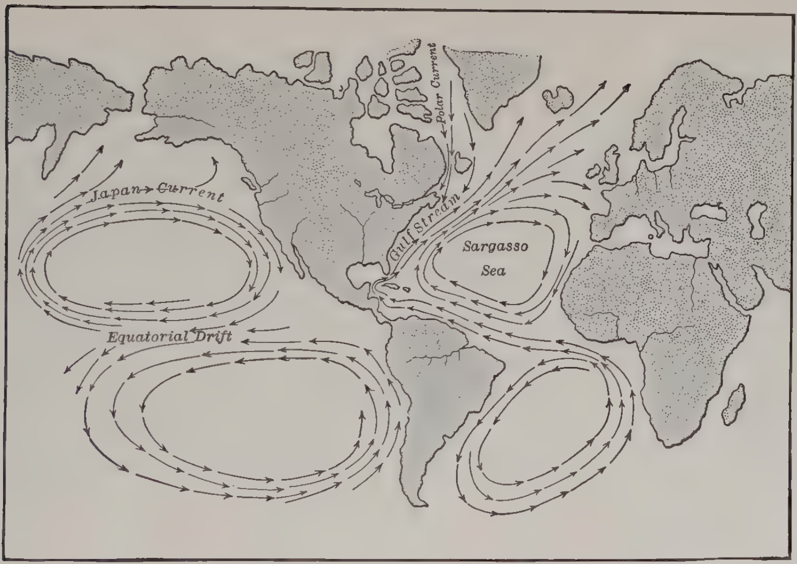


Fig. 124. — Map showing main ocean currents and drifts.

lantic, the equatorial current striking the north coast of South America is deflected northward. A part enters the Caribbean Sea and the Gulf of Mexico, whence it issues through the greatly confined Straits of Florida and passes northeast into the Atlantic as the well-known Gulf Stream. In the straits this current averages 72 miles a day and during the summer and winter months sometimes rises to 120 miles a day; but as it spreads and approaches mid-ocean the velocity diminishes greatly, falling finally to 10 miles a day.

As this drift approaches the shores of Europe, it divides and one part turns southward to pass along the coast of Africa, and so to rejoin the westward equatorial drift. Another portion passes northward into the Arctic Ocean, and to balance this a cold current comes down from the western coast of Greenland, past Nova Scotia and New England, and gradually passes under the warm surface of the Gulf Stream to sink into the abyss.

In a similar way, in the North Pacific a current moves along the coast of Asia, then eastward and finally southward along the western coast of North America. It is known as the Japan current. These warm currents, moving into high latitudes, have a great effect upon climatic conditions in the lands whose shores they strike. Thus the ocean currents, like the atmosphere, by taking part in the general circulation on the surface of the globe are great distributors of heat. They carry warmth into the polar seas and, returning as cold currents, bring with them reduced temperatures and icebergs to be melted in warmer regions. Were it not for their agency, ice would continually increase in the polar areas. The indirect effect of the ocean currents on geological processes is therefore important, although these currents perform little direct geological work.

Differences in temperature may generate ocean currents, the denser, colder water tending to move toward the lowest places and to displace the warmer water toward the surface. For this reason there is a slow and general movement of the polar waters along the bottom into all the deeper ocean basins, which consequently have a temperature approximately at the freezing point of surface fresh water. If there were no cold polar regions, the ocean bottoms would not be so cold.

Marine Erosion

Erosive Processes. — The surface of the seas, like an ever-moving horizontal saw, is ceaselessly cutting and gnawing away at the lands. This it does in several ways, producing a variety of features that are worthy of consideration.

Sea water has more or less solvent effect on rocks, tending to disintegrate them, and thus aid in their destruction; but the chief factor in marine erosion is the mechanical attack of the waves. The force of impact of large waves due to the sheer weight of tons upon tons of surging water may be sufficient to erode unconsolidated sediments rapidly or to move enormous blocks of stone. Dynamometer measurements on the coast of Scotland showed the average force of the summer waves to be 611 pounds per square foot and the average for the winter months to be 2086 pounds with extremes up to 6083 pounds. The damage done by storms to harbors and breakwaters bears further testimony to the force of the waves. During a great storm at Wick, Scotland, in 1872, a solid mass of stone and concrete weighing 1350 tons was torn from its place at the end of a breakwater and dropped unbroken inside the pier. When great waves strike against a sea wall or solid cliff, they not uncommonly dash up to heights of 100 feet or more. At the lighthouse on Tillamook Rock, on the exposed coast of Oregon, the water of the waves was thrown over 200 feet above the sea during a storm in 1902, and during the winters of both 1912 and 1913 the impact of the waves broke panes of plate glass in the lantern of the same lighthouse at a height of 132 feet above the sea. It is not surprising, therefore, that the faces of bold cliffs are shattered, and softer materials rapidly dislodged, simply by the impact of the waves. (Fig. 125.)

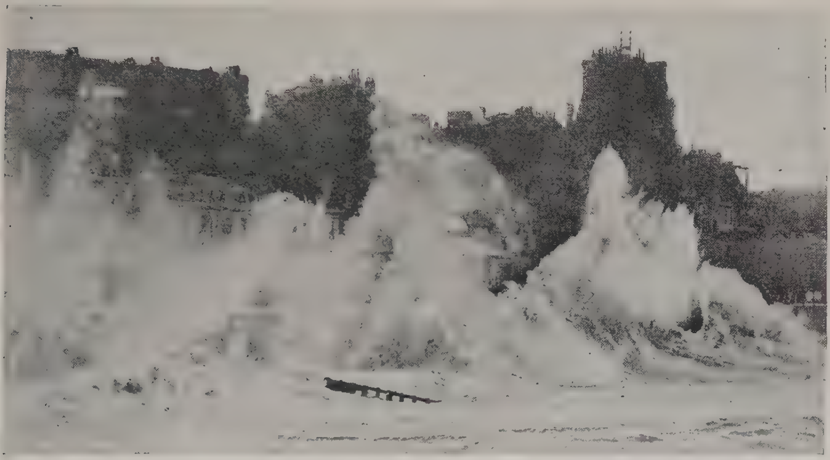


Fig. 125. — Storm waves breaking against the sea wall at Hastings, England. (After Johnson. Photo by Judges.)

Hydraulic pressure may be a powerful agent of the waves as well, for most rock masses have crevices or larger cracks in them, and the air or water in these, driven violently in by the impact of the waves, acts as a wedge, disrupting them and dislodging large pieces. In this way heavy masonry is often torn asunder. Moreover the water, rushing into cavities and suddenly retreating, leaves a partial vacuum which tends to suck away portions of the roof and sides, and the constant repetition of this action aids in forming sea caves, blowing holes, and spouting rocks, so frequently seen on rocky coasts.

The chief eroding action, however, is accomplished by abrasion, and in performing this work the waves use as tools the dislodged material, and also that which falls from above and tends to form a talus at the foot of the sea cliff. The constant striking and grinding, not only of sand and gravel, but even of heavy boulders, render the waves formidable agents of destruction, through whose work even the hardest rocks are worn away. The ineffectiveness of clear water, aside from the mechanically disrupting processes, is strikingly shown in places on the coast of Norway, where headlands extend into water too deep to be affected by coastal débris. The rock surfaces, smoothed and furrowed by former action of glacial ice, still retain these characteristic features, though subjected to the constant washing of the waves.

In the process of abrasion the material used by the waves is itself ground up and reduced to sand and silt. It loses its angular character and takes on the rounded form characteristic of coastal débris submitted to chafing by the waves.

Transportation of Sediment. — If wave erosion is to continue, the ground-up rock débris must be removed, like sawdust from the track of the saw, in order that fresh rock surfaces may be exposed to attack; otherwise the fine material would act as a buffer to receive the waves and prevent further erosion. This removal is done by the undertow, which takes the débris out to sea, and also by tidal and littoral currents, which carry it away. As in the case of streams, transportation takes place either in *solution*, in *suspension*, or by *bottom rolling*. The currents in the sea are aided greatly by the waves. When the particle of sediment is lifted free of the bottom, even the gentlest current can deflect its fall; in this way slow currents, aided by the repeated lift of the waves, can sweep along sediments that they alone would be powerless to move.

When waves reach shallow water and begin to drag heavily on the bottom, their movement ceases to be one strictly of oscillation, for the water then tends to roll ahead as a wave of translation which, like a strong current, drags the bottom material forward with it. Since the waves drag bottom only when approaching the shore, they tend to transport the bottom sediment landward. This tendency is opposed, however, by the undertow and by gravity, both tending to shift the sediment down the seaward slope. These seaward forces act continuously and are able, therefore, to overcome the stronger but intermittent surges of the waves. Although the sediment is dragged back and forth with each passing wave, the net movement of the fine material is seaward. On the other hand, the undertow may be unable to transport coarse material which the breaking waves drag forward, and when this is true the coarse sediment migrates landward even though the fine material is being shifted seaward. Murray has recorded, for example, that stone ballast discharged from ships near the British coast in water as much as 60 feet deep, has been thrown up on the beach during a subsequent storm. This factor keeps the coarse sediment concentrated at the shore, where it is known as the *beach shingle.*

Even fine sand is thrown shoreward to form a sandy beach where loose sediment is sufficiently abundant, the waves carrying forward more of the material than the undertow can remove. The beach is but a temporary accumulation of sediment in transit, however, for there is a constant loss to the undertow and shore currents that spread the material seaward.

Physiographic Development of the Shore

In their attack on the shore the waves and currents carve erosive features in some places and in others build up deposits of the loose sediment. The nature of the shore features thus produced at a given place depends partly upon the character of the land surface against which the marine agents have to operate and partly upon the length of time they have been at work. As a land surface under the influence of stream erosion passes through a cycle of physiographic changes from youth to maturity and old age, so also the shoreline passes through a predictable series of changes as the marine cycle progresses, and many of the shore features, like those of a degrading land surface, are temporary features in the physiographic cycle.

The marine cycle is inaugurated ordinarily in one of two ways; either (1) by an uplift of the sea floor against which the waves start their attack or (2) by a submergence of the coast which brings the sea into contact with a former land surface. The initial form of the shore on an emergent sea floor is utterly different from that on submerged land, and the sequential forms developed from the one are so distinct from those of the other that each type of shore must be described separately.

The Shoreline of Emergence. — When a sea floor emerges by gentle uplift it forms a nearly flat *coastal plain* mantled by unconsolidated sediments. The plan of this shoreline is at first very simple, since it marks the intersection of this flat surface by sea level. The water is so shallow for a considerable distance from shore that the waves drag heavily upon the bottom, picking up the loose sand and gravel, which they carry forward to the line of breakers. Here the loose material is dropped again as the wave spends itself, gradually building up a narrow submarine ridge parallel to the shore and just within the line of breakers. Storm waves eventually build the deposit above sea level, forming a long low sandy island parallel to shore (Fig. 126, *A*). This structure is known as an *offshore bar* or *barrier beach.* The narrow body of shallow water lying between the offshore bar and the shore is a *lagoon.* Large lagoons are also called *sounds* but not all sounds are lagoons.

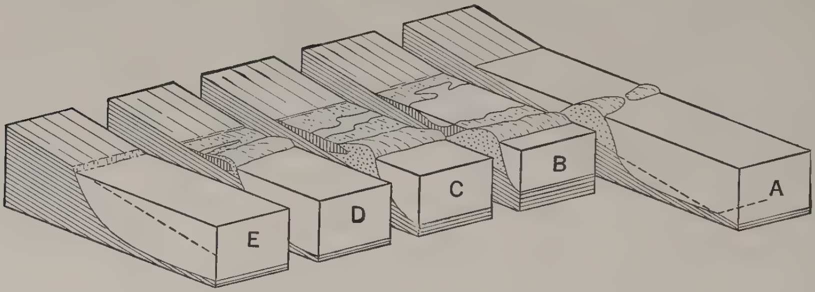


Fig. 126. — Showing youthful stages in the development of a shoreline of emergence. The offshore bar is stippled in section and the lagoon filling is vertically lined. *A*, early youth, showing the formation of the offshore bar; *B*, a later stage showing partial filling of the lagoon; *C*, a still later stage in which the offshore bar has been prograded toward the shore and the lagoon has become a fresh-water marsh; *D*, late youth; *E*, early maturity, the offshore bar and lagoonal deposits having disappeared. Vertical scale exaggerated. (After Davis.)

Ordinarily the sweep of the tide in and out of the lagoon keeps inlets through the barrier open to the sea. The Atlantic and Gulf coasts of the United States display these features in grand development (Fig. 137). The width of the lagoon depends on the seaward slope of the bottom and the size of the storm waves. It is commonly a mile or more and may be several miles. At Cape Hatteras the offshore bar is about 20 miles from the mainland but along the east coast of Florida it is near shore and the lagoon is known as Indian River. The city of Galveston is built on an offshore bar and its tragic flood of 1900 occurred when high seas, driven by a hurricane, rose 15 feet above their normal level.

The sea floor outside the barrier is gradually excavated by the waves, supplying the material of which the barrier is made. As its depth increases the waves are less impeded by friction on the bottom and are finally able to break with force against the barrier itself, eroding material from its seaward side. A part of this débris is carried out to sea, and during storms a part of it is thrown over the bar into the edge of the lagoon. In this way the barrier gradually moves shoreward. Meanwhile the lagoon, rarely more than 20 feet deep at the start, tends to be filled by sediment washed into it from the land, as well as by that thrown over the barrier (Fig. 126, *B*). Where the water is shallow enough, vegetation thrives and adds its quota to the accumulating sediment. As the barrier is gradually driven shoreward the lagoon becomes a salt-water marsh with open tidal channels (Fig. 126, *C* and *D*). Eventually both barrier and lagoonal deposits are completely cut away as the sea bottom is excavated up to the shoreline (Fig. 126, *E*) and then for the first time the waves begin their attack upon the land.

During these early stages of the cycle, the bottom profile near shore has been gradually changing from the simple slope of the original sea floor to the compound curve represented in Fig. 127. There is a concave portion near shore where the bottom is being eroded. It extends seaward as a nearly flat *wave-cut bench* or *terrace* to the line of depth where wave cutting is ineffective. The depth of the wave-cut terrace depends on the size of the prevailing storm waves and the supply of sediment they have to move, but it is generally between 10 and 20 feet near shore and gradually deepens seaward through abrasion by the sediment that is shifted across it. Beyond the zone of cutting the sediment is dropped and gradually built out as a submarine embankment or *wave-built terrace*, which continues the nearly flat slope of the wave-cut terrace. The bottom is convex over the outer part of this deposit and concave further out where the sea floor drops away to the normal bottom slope. The curve shown in Fig. 127 is the *profile of equilibrium* and its general character will be maintained to the end of the marine cycle, though the curve will gradually flatten out as the sea cuts farther inland and the

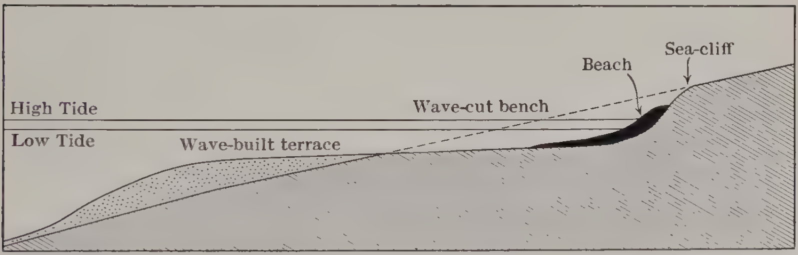


Fig. 127. — The wave-form shore profile of equilibrium, concave near the shore where the sea is cutting and convex farther out where sediments are accumulating. The beach is represented by solid black and the sediments of the wave-built terrace are stippled. Vertical scale exaggerated.

waves waste more and more of tneir energy in bottom friction while crossing the wide neritic zone.

Where the waves are cutting into solid rock the wave-cut terrace is rock-floored, though it may be more or less mantled with débris in transit from the shore; but where the attack is upon unconsolidated sediments the wave-cut terrace obviously has a floor of loose material.

As the waves concentrate their attack at sea level, cutting horizontally into the sloping land surface, they tend to undercut the land. The overhanging material is gradually dislodged and falls down and thus the coast comes to be terminated in a *sea cliff* as in Fig. 126, *E*. The nature of the sea cliff depends in part on the material attacked by the waves. In firm rocks it may be very steep or even overhanging, but in loose material like sand, slumping and sliding keeps pace with undermining and the sea cliff is a slope equal to the angle of repose of the loose material. The form of the sea cliff is also influenced by the relative rate at which the shore material yields to weathering and to wave cutting. For example, clay and chalk are easily cut by the waves and commonly present bold bluffs in spite of their softness, whereas hard granite may succumb more rapidly to weathering than to the attack of the waves, especially where it is much jointed and exposed to frostwork. The hard rock may thus present a sea front of low slope whereas soft rock may form cliffs.

The débris that falls from the sea cliff or that is eroded at sea level (as well as that which is introduced by streams) is shifted back and forth by the waves and currents and the finer material is carried seaward with the undertow. The coarser sediment is kept at the shoreline, however, to form the *beach*, a low ridge of sediment thrown up by the waves along the shore. As indicated in Fig. 128, the beach is only a thin veneer on the bedrock floor, extending landward to the limit reached by the greatest storm waves. In cross section it is convex upward at the shore, where it is thickest. It extends for a short distance under the water, thinning gradually and merging into the sheet of sediment that partly mantles the rock-cut bench while in transit out to sea. The upper margin of the beach is generally marked by a belt of coarser material thrown up by the heaviest storm waves and this grades into finer material toward its seaward margin. In places where the shore zone is formed wholly of fine sediment the beach is made of sand alone.

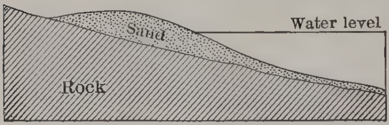


Fig. 128. — Section of a beach. (After Gilbert.)

If the rock forming the sea cliff is cut by vertical joints, the waves tend to widen the joints and quarry out some of the blocks, leaving others standing isolated as *stacks* (Fig. 129). If the joints are more irregular and intersect before reaching the summit of the cliff, *sea caves* are formed at the base of the cliff. As the waves break into the mouth of a sea cave they exert a heavy hydraulic pressure on its sides and roof, followed by rarefaction of the air as the wave recedes. By this action the roof of the cave may be excavated through to the surface so that spray will dash up through the roof with each breaking wave. Such caves are known as *blow holes* or *spouting caves.* Some narrow promontories or islands are undercut by the development of sea caves to form *sea arches.*

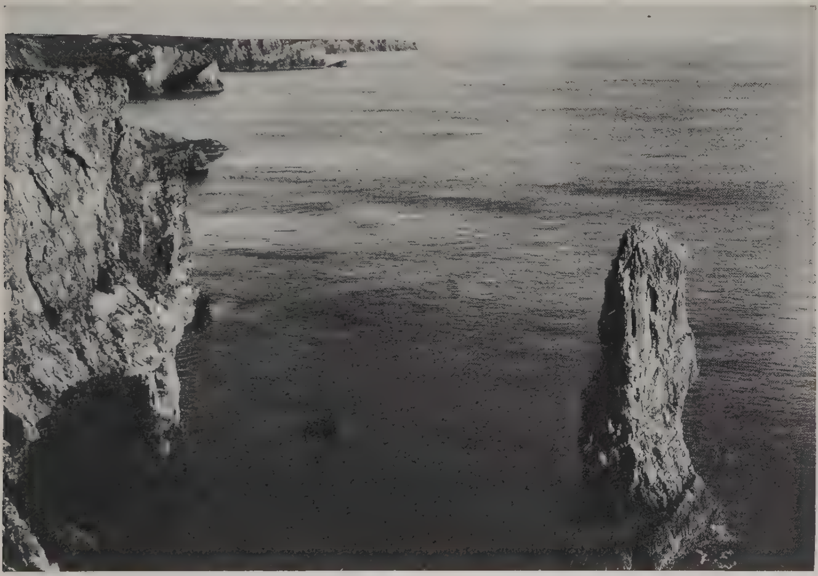


Fig. 129. — Sea cliff and stack, the latter a remnant of former land now eroded away. Coast of Wales. (Geol. Surv. of England and Wales.)

Obviously the greatest inroads are made by the sea where the rocks are weak and the slightest progress where they are resistant. For this reason a simple shoreline formed of heterogeneous rocks gradually develops irregularities as erosion progresses; the resistant masses stand out to form *headlands* or *promontories* and the weak rocks recede to form *coves* or *bays* (Fig. 130). If the rocks are composed of parallel layers or beds and their edges are exposed to the waves, the weaker, softer layers are rapidly worn away and the harder beds, left unsup-

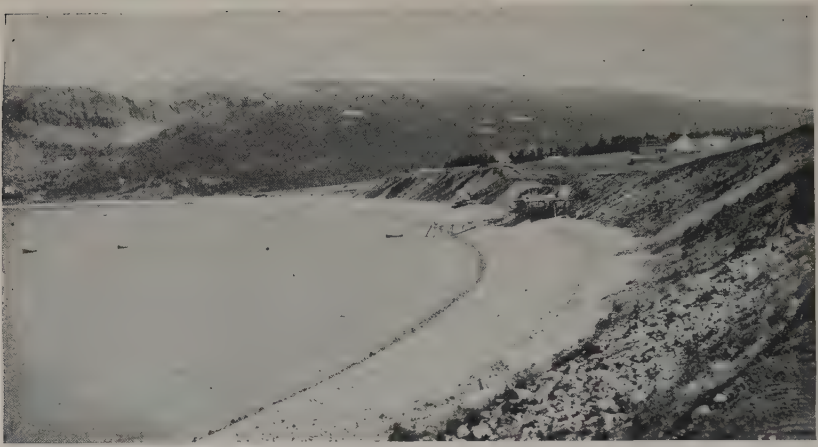


Fig. 130. — A bay with curving beach. Conception Bay, Newfoundland. (Walcott.)

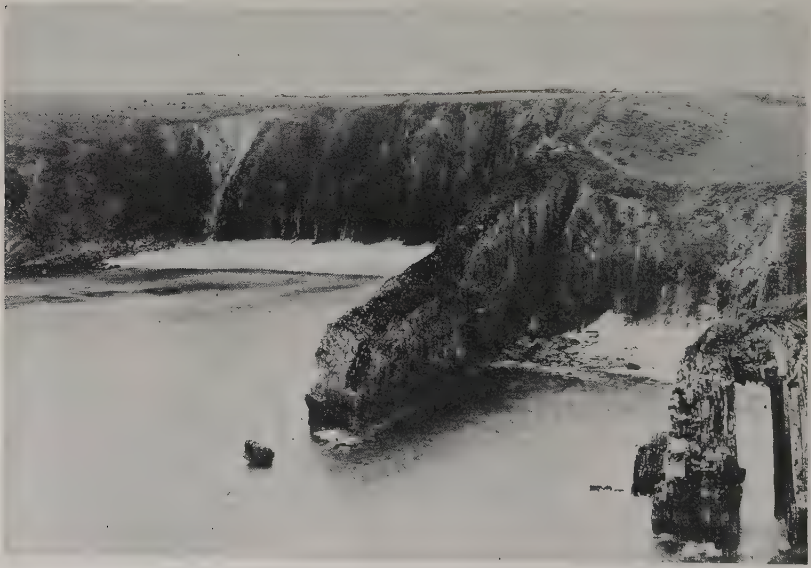


Fig. 131. — Irregular coast and sea cliff produced by erosion in nearly vertical rock strata Pembroke, Wales. (Geol. Surv. of England and Wales.)

ported, break away in blocks. If the beds are horizontal the harder layers may project for a time as table rocks with cavities under them, as on the coast of Lake Superior. If the beds are vertical, or inclined, but with edges exposed, the hard layers stand out like columns or ribs (Fig. 131). If the face of the beds is towards the sea, erosion is slower because the hard layers form an apron, or wall, to protect the soft layers behind them. If the rock masses are homogeneous and hard, like trap or granite, the irregularities are largely determined by joints and by the arrangement of these with respect to the sea front. Thus a bold coast facing the sea is likely to show many minor irregularities of topography.

In the physiographic development of the shoreline of emergence, the stages illustrated in Fig. 126 are all preliminary to the real attack against the land. They represent the *youthful stage* of the marine cycle. Following the disappearance of the offshore bar, the sea launches its drive with full vigor against the shore. This is the beginning of the *stage of maturity.* The sea cliff is now formed and the simple shoreline rapidly develops irregularities, until the plan of the shore is adjusted to the forces of attack and to the resistance of the rocks. Erosion then proceeds gradually along the whole line. The final stage of *old age* is attained only when the sea has cut so far inland that the waves spend most of their force in friction over the shallow bottom and make but a feeble attack upon the shore. After this stage is reached, further retreat of the shore due to marine erosion proceeds very slowly.

The Shoreline of Submergence. — The initial form of a shoreline of submergence is marked by extreme irregularity, since the marine water comes to rest against an eroded land surface, entering far up the valleys and forming deep irregular embayments between sloping headlands. Canyon-like glaciated valleys thus drowned become *fiords,* and the lower courses of the normal stream valleys become *estuaries* or deep bays. Hard-rock islands are likely to be abundant near the shoreline, representing isolated hills of the drowned land surface.

The *youthful stage* of the cycle of erosion of such a shoreline is marked by lack of adjustment of the erosive forces to the plan of the shoreline. The waves must concentrate their attack chiefly on the offshore islands and the headlands that defend the deeper embayments. These are quickly cliffed, while the bay-heads tend to be silted up by the streams that enter them. In their fresh attack against the land surface, the waves quickly indent the weaker spots, developing *sea caves, arches,* and *stacks* and forming little bays even in the headlands. The irregularities thus developed are on a small scale and are only incidents in the general process of straightening the shoreline, but they give to it a crenulate plan characteristic of early youth.

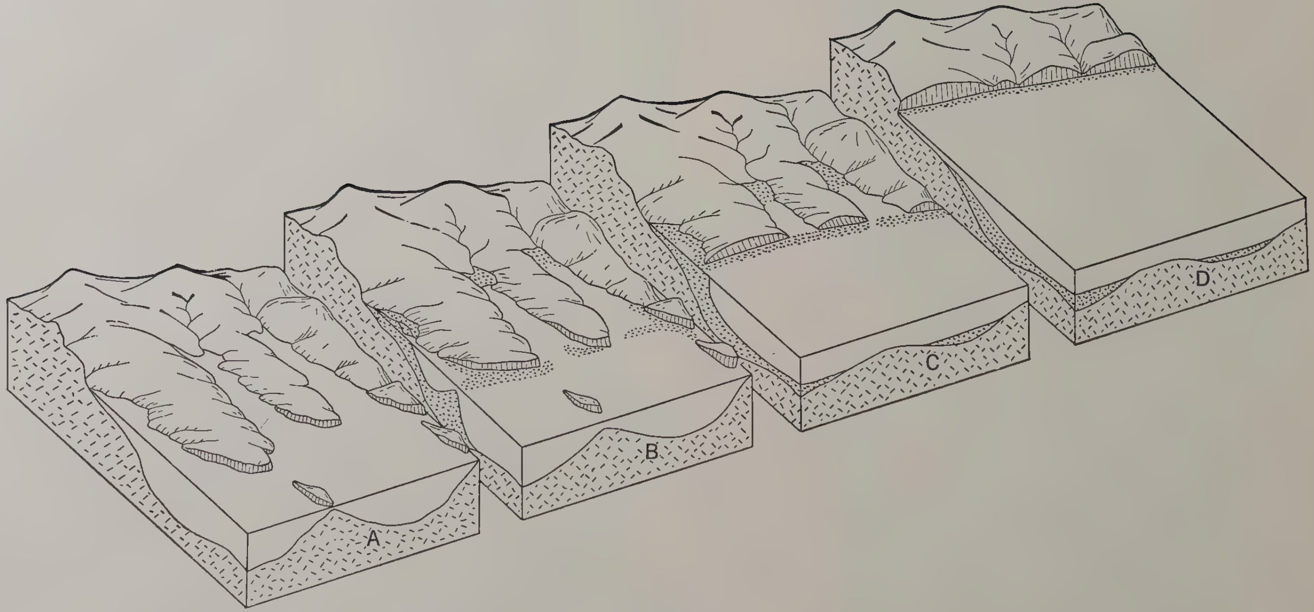


Fig. 132. — To show early stages in the development of a shoreline of submergence. *A*, early youth, showing cliffed islands and headlands without beaches; *B*, a later stage showing recession of the headlands and the development of winged headlands through the growth of spits and hooks; *C*, later youth showing truncated headlands united by bars and extensive salt-water marshes in the heads of the bays; *D*, early maturity. The scale of this figure is much smaller than that of Fig. 126, so that each block represents a larger area than in that diagram. (Modified from Johnson.)

Three youthful stages and one of early maturity of the submergent shoreline are indicated in Fig. 132. It will be noted that they tend toward a simplification of the extreme irregularities of the initial shore. The first stage is marked by *cliffed headlands.* Since the surface is drowned, deep water comes close to the headlands and the débris first eroded sinks beyond reach of the waves so that no beach is produced. Gradually the headlands are cut back and the débris supplied by their erosion accumulates until a deposit is built up into the zone of wave action. A beach is then formed at the foot of the cliff. Littoral currents do not enter deeply into the estuaries and as they sweep past the truncated ends of the exposed headlands they carry part of the beach material laterally into deeper water where it settles, building out a submarine embankment. By continued action of this process the embankment is built out across the adjacent embayment very much as an artificial fill is made across a valley by the dumping of dirt at the end of the fill. Incoming waves will throw some of this material upon the embankment and build it above sea level. It then appears as a low sandy or gravelly promontory built out from the beach. If nearly straight this structure is a *spit.* Another current crossing the end of a spit may sweep the material aside, developing a curved or hooked spit which is known as a *hook* (Fig. 133). If the deposit is extended until it closes, or nearly closes, the mouth of an embayment it is known as a *bar* (Fig. 134). Spits and hooks tend to form wherever the littoral currents

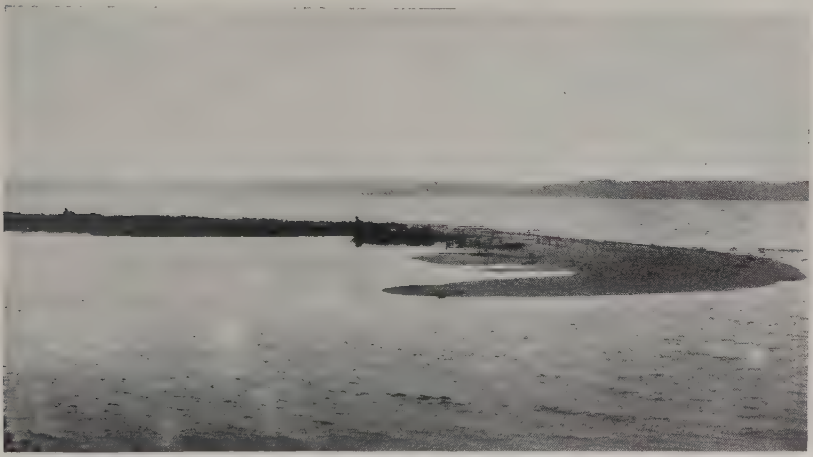


Fig. 133. — A curved spit, or hook. Duck Point, Grand Traverse Bay, Lake Michigan. The left end, beyond the edge of the view, is attached to a point of land. (U. S. Geol. Surv.)

sweep sediment away from the beach, whether it be at the end of a promontory or merely of a blunt cape. Bars tend also to form in the sheltered water between islands and the shore, resulting in *land-tied islands* known as *tombolos* (Fig. 135). *Sea caves, arches,* and *stacks* are especially characteristic of these early youthful stages of the submergent shoreline.

While spits, hooks, and bars are forming about the retreating headlands the inner ends of the embayments tend to be filled by deltas where streams enter these protected bodies of deep water.

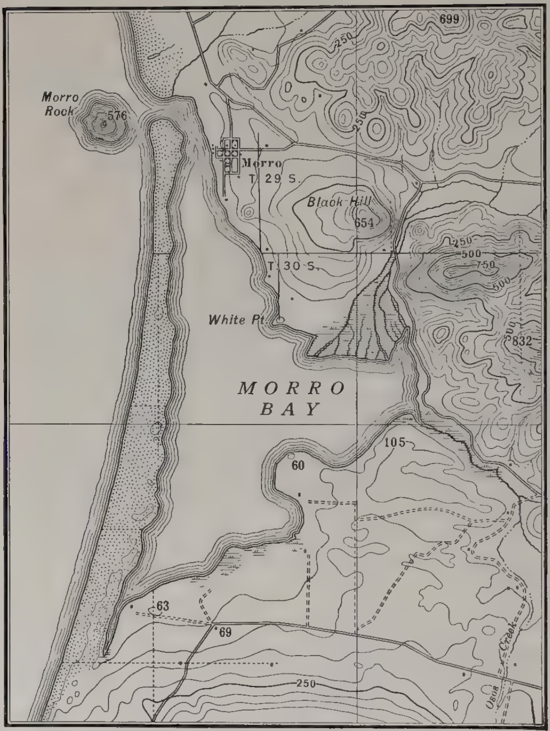


Fig. 134. — A bar closing Morro Bay, Calif. (U. S. Geol. Surv.)

Eventually, as in Fig. 132, *C*, the shoreline is much simplified as the land-tied islands are cut away and the headlands truncated and united by more or less continuous bars. The bay-heads become largely filled with sediment and pass into salt-water marshes like those that are so common along the New England coast at the present time.

When finally the headlands have been completely cut away and the bay-head fillings are all removed, as in Fig. 132, *D*, the *youthful stage* of submergent coast is completed and *maturity* has begun. The waves are now at work along the entire coast, the plan of the shoreline is adjusted to the resistance of its rock masses, and the bottom has developed the *profile of equilibrium,* concave over the wave-cut bench and convex over the wave-built terrace. The deeper irregularities in the sea floor have been filled with sediment. From this stage on the development is identical with that of the shoreline of emergence.

Shorelines Produced by Interrupted Marine Cycles. — Completion of the marine cycle would require stability of the land and a constant sea level for a very long time. A relative uplift or depression of the coast will interrupt the cycle and superpose a new cycle on the old.

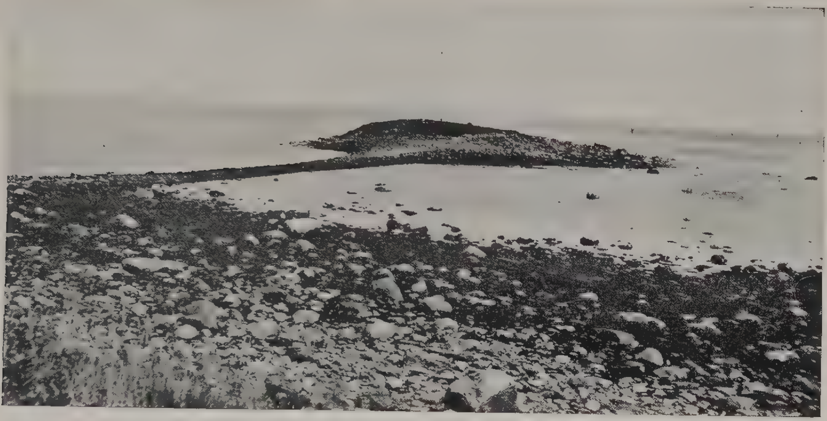


Fig. 135. — A tombolo or land-tied island. (U. S. Geol. Surv.)

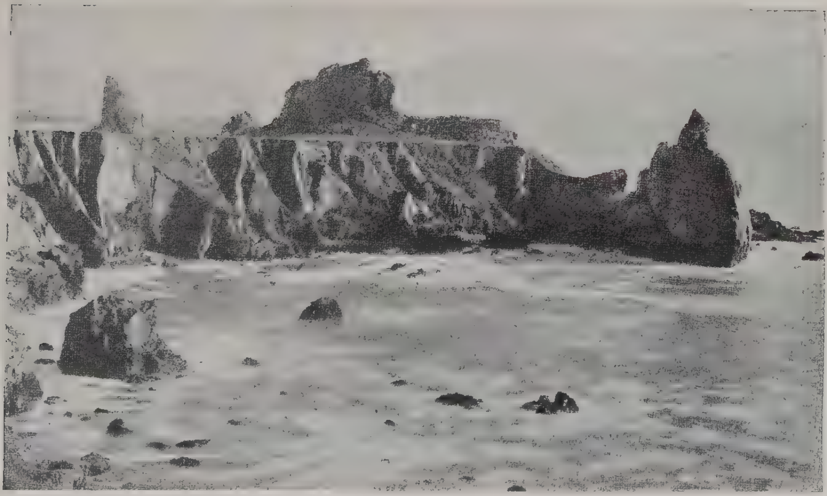


Fig. 136. — An elevated wave-cut terrace north of Port Harford, Calif. Old stacks rise above the terrace. (U. S. Geol. Surv.)

If uplift of the coast is intermittent, the wave-cut bench formed during a period of stationary sea level may be lifted above the sea to form a *marine terrace* paralleling the shore (Fig. 136). Such terraces are strikingly developed for long distances along the southwest coast of Newfoundland. At Port au Port, one of them bevels evenly across the edges of steeply tilted strata and attains a width of more than 2 miles. Although originally carved by the waves, it is now 55 feet above sea level. Abundant marine shells occur in the clay that mantles another terrace lying just 100 feet above sea level along the western end of Cape St. George, Newfoundland. Farther north in both Newfoundland and Labrador, there are remnants of older and more eroded marine terraces at several elevations up to about 500 feet above the present sea. Along the coast of California, marine terraces have been recognized up to 1600 feet above sea level.

The landward margins of some marine terraces are marked by ancient sea cliffs accompanied by other shore features such as stacks and sea caves. Characteristic beach deposits also mark the position of some old shorelines, and where uplift has been intermittent, a series of beaches may lie one above another as they do along the coast of southeastern Labrador.

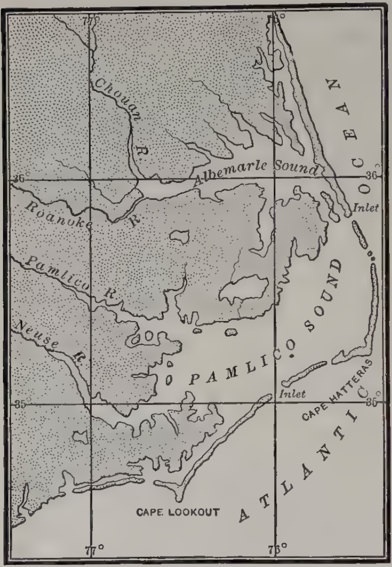


Fig. 137. — Map of the coast of North Carolina showing an association of drowned valleys, wide lagoons, and offshore bars.

Coasts frequently undergo uplift followed by submergence, or depression followed by later uplift. The result of such oscillations in the level of land and sea is the production of a variety of shore features some of which have resulted from emergence and others from submergence. Thus, for example, the Central Atlantic coast of the United States (Fig. 137) is marked by the drowned valleys characteristic of a submergent shoreline and by the offshore bars characteristic of an emergent coast. An extensive submergence has been followed by a recent but lesser emergence, leaving the valleys drowned and at the same time reducing the depth of the near-shore water to such an extent that barrier beaches have been developed. Under these conditions the barrier beach at first has extensive interruptions marked by the submerged valleys. The barrier beach may be extended across these depressions as a result of lateral shift of sediments by littoral currents, growth of the barrier beach taking place by the development of subsidiary spits and hooks and bars. There are good examples of this relationship along the coast of New Jersey.

Some coasts show old, dissected marine terraces far above sea level, bearing witness to an earlier uplift of the land, whereas the present shoreline is indented by drowned valleys as a result of more recent submergence.

Distribution of Marine Sediments

Shallow-Water Terrigenous Deposits. — The land-derived (terrigenous) marine sediments all have as their goal the quiet depths of the sea or ocean bottom, their temporary deposition in the form of beaches, bars, and barriers being merely an incident in their seaward journey. Constant agitation by the waves during their transportation imparts to the marine deposits a much higher degree of sorting than is usually attained by river-laid sediments, a fact suggested by the cleanness of fine beach sand.

Where wave and current action is vigorous, only coarse material can accumulate, but as this activity declines, finer and finer sediment comes to rest. The coarse sediments are therefore deposited in shallow water, and, in general, the deposits grow finer of grain with increasing depth. But since very shallow sea bottoms may be protected by sheltering land masses or by coral reefs, it must not be supposed that fine-grained sediments are all of deep water.

In the neritic zone of the sea the sediments vary greatly in character from place to place. For example, along the bold and exposed coast of California the near-shore sediments include gravel and coarse sand; off the mouth of the Mississippi the deposits are finer sand and mud derived from distant highlands; and the low limestone coast of southern Florida is bordered directly by deposits of the finest limy ooze.

Soundings show that modern gravels are generally confined to depths of less than 30 feet and are deposited within 1 mile of shore, though exceptionally gravel is carried to depths of 150 feet and to distances up to 10 miles from shore. Sand generally does not extend beyond a depth of 250 or 300 feet and the outer portion of the continental shelf is therefore covered by muds alone.

Coral Reefs and Calcareous Deposits. — In the shallow water along many tropical and subtropical coasts coral reefs play an important geological rôle. These limy reefs are estimated to have a combined area of half a million square miles at the present time and the calcareous débris produced through their erosion by the waves is spread over a much greater area of the sea and ocean floors.

Corals are small marine animals characterized, among other things, by the fact that they secrete an external limy skeleton about the base of the body. The name *stony coral* is commonly applied to the skeleton in distinction from the living animals which are called *polyps.* The polyps bud from the margin of the parent in plant-like fashion and retain a fleshy connection, so that living colonies of many thousands of individuals combine in the secretion of an elaborate limy structure of considerable size. Some of the colonies are massive and hemispherical or irregular in their growth, whereas others are branching and shrublike. Some individual colonies reach a diameter of 15 feet or over and house a million or more little polyps, each living in its own tiny protecting chamber like a cliff-dweller in a city of stone (Fig. 138).

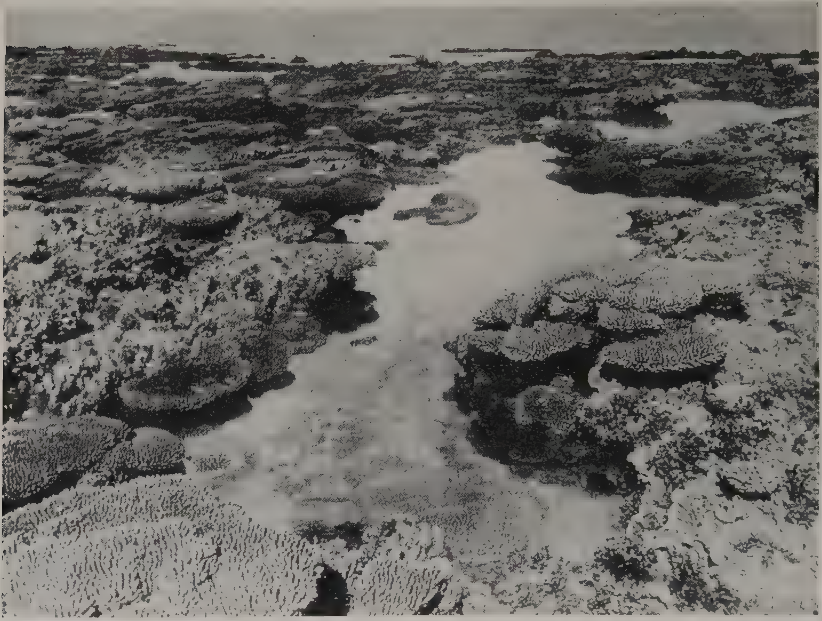


Fig. 138. — Growing corals, seen at low tide. The view is from a barrier reef across the lagoon toward mainland 20 miles away. Great Barrier Reef, Australia. (Saville Kent.)

As the corals expand and build upward the older polyps die, and gradually the dead base is buried by the growth of other animals and by the abundant débris which storm waves break from the living parts above. This accumulating deposit is eventually cemented into a white cavernous limestone, upon whose upper surface the living corals grow and flourish. In this way the *coral reef* is formed. It rises until its surface is just below sea level, where it is laid bare only at the lowest tides for short periods, for corals can stand exposure to the air but a few hours at a time. Over the reef the waves swirl and break, and at its outer edges the reef-building corals thrive best, for in the rush and dash of the waves they find the most food and calcium carbonate in the water, and the richest supply of life-giving oxygen.

Coral reefs are sharply limited in their distribution by the depth and the temperature of the water. Although individual corals of various kinds live in more varied environments, the reef-forming kinds thrive only where the water is clear, normally saline, shallow, and warm. They cannot endure a temperature below 68° F. and they live normally where the water is less than 150 feet deep, though rarely, where there are descending warm currents, they exist well below this limit. Coral reefs are therefore confined to the shallow waters of the tropics and near tropics, except where (as at Bermuda) warm currents form exceptional conditions. The distribution of fossil coral reefs in the older geologic deposits is consequently thought to carry special climatic significance.

According to their position and form, coral reefs have been grouped into three general classes: fringing reefs, barrier reefs, and atolls.

Fringing reefs lie close against the shore and form a bench or platform extending out toward the sea, laid bare only at very low tide. As the corals cannot grow above sea level the growth of this platform is chiefly seaward. Opposite the mouths of streams the reefs are nearly always wanting because the corals cannot endure the freshened and muddy water.

Barrier reefs differ from those just described in that they are situated some distance from the land, and are separated from the shore by a *lagoon* of shallow water. Many of the high volcanic islands of the Pacific as well as the islands of the Caribbean region are more or less completely girdled by such encircling reefs. The west coast of the island of New Caledonia has a reef of this sort that extends for 400 miles, and the greatest of all coral reefs is the Great Barrier Reef which stretches for 1200 miles along the eastern coast of Australia, with an average distance of 20 to 30 miles from shore and a depth of 100 to 300 feet of water in its great lagoon.

Atolls are more or less ring-shaped reefs enclosing circular lagoons instead of islands (Fig. 139). The breadth of such rings may be but a fraction of a mile or it may be as much as 20 or even 50 miles, and the depth of water in the lagoons ranges from a few feet up to 300 feet. On the outside some of the slopes descend quite sharply thousands of feet to the ocean floor. Generally there are openings in the reef on the leeward side, which afford access to the lagoon.

Atolls, like either of the other types of reefs, become low islands where storm waves toss the coral débris upon the reef and build it above sea level. The islands thus made are usually not more than 10 or 15 feet above normal sea level and from a quarter to half a mile wide, though many of them are long in the direction of the reef. Most of them are covered by vegetation, including palms, and form spots of great beauty. Some are inhabited, though their low elevation subjects them to the danger of being swept by the sea during exceptional storms. The only living atoll in America is small Sombrero of the West Indies.

The evidence indicates that atolls represent a fringe of coral growth around the edges of subsiding volcanic islands.

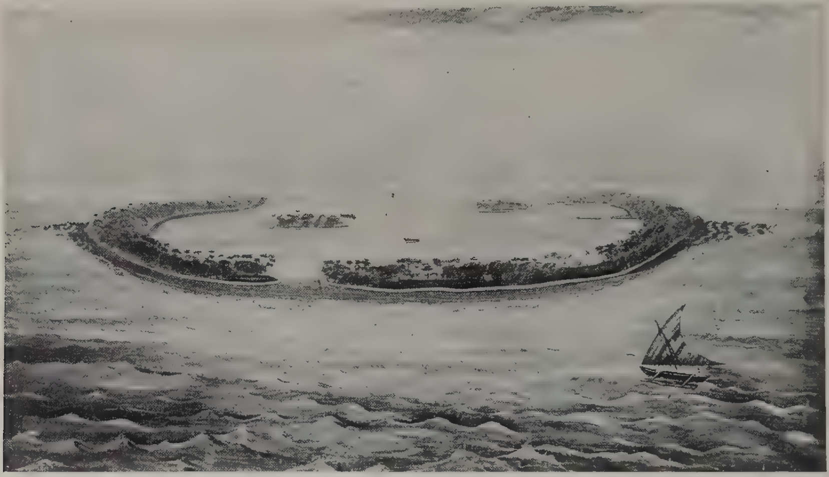


Fig. 139. — An atoll. (After Dana, from an old picture.)

Oozes of the Ocean Bottom. — The abyssal ocean floor, constituting three-quarters of the Earth's surface, is covered to an unknown depth by exceedingly fine-grained, soft deposits known as oozes. These sediments are of several kinds and of different sources.

Red clay is the most extensive of the oozes since it mantles the deepest parts of the ocean floor to an extent of over 50,000,000 square miles, an area about equal to the entire land surface of the Earth. Its sources are believed to be windblown dust (derived both from erosion of the land and from volcanic eruptions), pumice that has floated for a time before sinking, meteors and meteoric dust that have fallen directly into the oceans, and also the insoluble residues of organic structures that have sunk to the ocean bottom. All of this material is so thoroughly altered by chemical decay that the original source material for any sample of the clay is obscure. The red color of this extensive deposit is probably the result of the extreme slowness of its accumulation, which gives time for its thorough oxidation in spite of the deep covering of marine water.

The remaining abyssal deposits are the *organic oozes* formed largely of minute shells or shell fragments dropped by organisms that live in the surface waters of the ocean. Some of these deposits are made of limy shells, and others are formed by the accumulation of microscopic siliceous shells.

Several distinct types of organic ooze are recognized, each being named for the group of organisms most important in its formation. The most widespread of the limy oozes is that known as *globigerina ooze* (Fig. 140). It is formed largely of the microscopic shells of single-celled animals known as Foraminifera. There are about twenty species of these animals that live floating in incredible numbers at the surface of the warmer oceans. In the reproduction of these tiny creatures the mother forsakes her shell and subdivides into a great many tiny daughters, each of which in turn secretes a new shell as it grows up, only to repeat the process. As reproduction is rapid and generation succeeds generation quickly in such forms, their abandoned shells drift down like a perpetual snowfall into the depths. Wherever these little shells fall below a depth of about 15,000 feet, the carbonic acid concentrated in the cold bottom waters again dissolves the shells, so that globigerina ooze is not found in the deeps of the ocean, but on the shallower parts of the abyss it mantles a vast expanse of approximately 30,000,000 square miles.

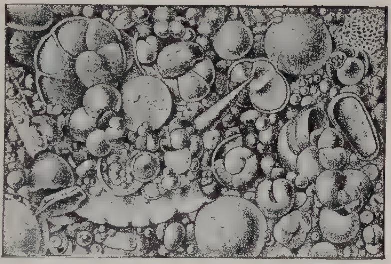


Fig. 140. — Globigerina ooze, much magnified. (After Agassiz and Murray.)

Radiolarian ooze is formed by the Radiolaria, another group of floating microscopic animals which, unlike the Foraminifera, fashion their delicate and beautifully ornate shells from silica. *Diatom ooze* is likewise formed by a group of microscopic plants, the diatoms, which secrete capsule-like shells of silica.

The oozes of the abyssal region are distinctive, and the general rarity of this type of deposit in the sedimentary rocks of the lands is believed to indicate that the continents have always remained continents and with very limited exceptions have never been covered by abyssal depths of water.

CHAPTER IX

SEDIMENTARY ROCKS

Sedimentation and Stratification. — The muddy water of swollen streams is turbid with sediment in transit toward the sea. Some of this material comes to rest in lowland areas where the streams spread in flood, but the final goal for most of it is the sea floor. In either case the sediment is spread in layers of mud or sand, and much of it is compressed and cemented eventually into *sedimentary rocks.*

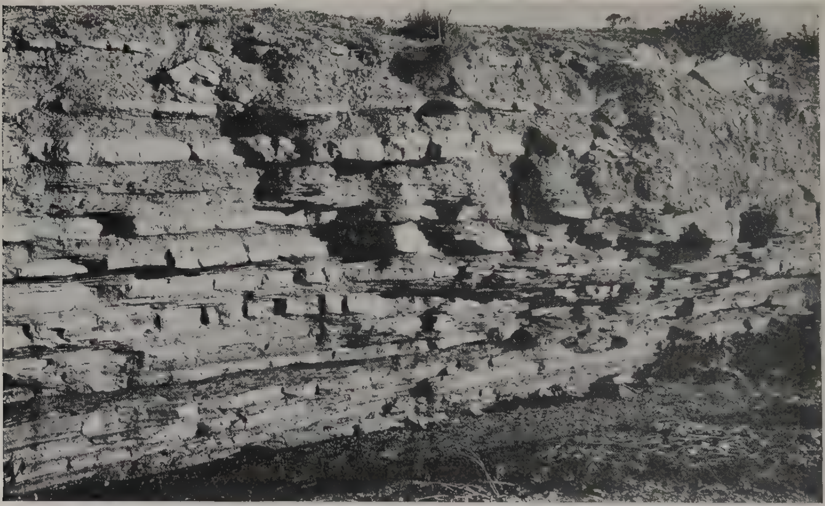


Fig. 141. — Regularly bedded sandstones and shales, near Pueblo, Colo. (U. S. Geol. Surv.)

The layered or stratified nature of such deposits arises from an intermittent supply of sediment or from changes in the velocity of the currents. If sediments of various degrees of fineness were dropped together into still water, the heaviest and coarsest would reach bottom first and upon them would settle the next in size and so on up to the very finest. There would, of course, be no distinct layers but a complete gradation from bottom to top. But if a second lot of sediment were introduced after the first had settled, it would form a distinct layer with its coarse base resting on the fine-grained portion of the first layer. In similar fashion the sediments introduced into a lake or the sea after one storm will spread and settle before the next storm or flood.

If, on the other hand, a mass of heterogeneous sediment is dropped into a steady current of moving water, the gradation takes place in a horizontal as well as a vertical direction, the successively finer material being dropped farther and farther along the bottom, or, in other words, being graded but not stratified. If the velocity of the current is altered and a second supply of sediment introduced, the new deposit will be evenly graded like the first, but its coarseness at any point will not correspond exactly with that of the first layer immediately below it. The two layers will be separated therefore by a distinct *bedding plane*, on either side of which they differ in texture. Since the velocity of streams and of the shallow marine currents changes from day to day and varies from place to place, it is inevitable that the sediments they transport will be deposited in parallel layers that differ in thickness, texture, and materials. Each layer is known as a *stratum*, the bedded or layered deposits are said to be *stratified*, and the condition is known as *stratification* (Fig. 141).

Source and Nature of Sediments. — It has been shown in previous chapters how each of the erosive processes contributes its share to the endless stream of rock waste which the rivers move toward the lowlands and the sea. This land waste varies in coarseness from the great boulders moved by glaciers or torrential streams to the finest of mud; and it also includes material which, like salt and calcium carbonate, is borne in solution. At the same time, it varies in composition from bits of fresh and unweathered rock to the end products of chemical decay. It is roughly classified, chiefly on the basis of coarseness, into *gravel*, *sand*, *silt*, and *clay*.

Gravel is an aggregate of coarse sediment in which individual particles have a diameter of 2 mm. or more. The smaller pieces, ranging in size from that of grape seed to that of baseballs, are known as *pebbles*, and the larger ones as *boulders*. Besides the pebbles and boulders, gravels usually include more or less sand.

The fragments that make up the gravel are always more or less water worn. At their source they were irregular, angular pieces of rock bounded by joints or fracture planes, but in being rolled along downstream, or tossed back and forth by the waves, their angles and corners suffer the most abrasion and the pebbles and boulders become more and more perfectly rounded. Rough and subangular fragments, therefore, indicate that they have suffered little transportation and that they do not lie far from their place of origin.

In their transportation, the pebbles and boulders of softer material are the first to be destroyed by reduction to sand and gravel. Since quartz is the hardest of the common rock-forming minerals it is the predominant substance in most gravels; and well rounded pebbles and boulders, having suffered long wear, are usually composed almost entirely of quartz. On the other hand, coarse sediment that has not travelled far commonly includes other minerals and rocks, such as granite, schist, basalt, or limestone.

Sands are sediments composed of grains smaller than gravel, yet exceeding 1/16 mm. (i.e., about the thickness of this page) in diameter. Such material is coarse enough so that it will not form a coherent, plastic mass when wet. Ordinary sand is like granulated sugar in fineness. With a lens it may be seen that the coarse sand grains are rounded like pebbles by attrition, but the finest grains are usually angular. This does not mean that the larger grains have travelled farthest but rather that grains below a certain size are not rounded in transit, the explanation lying in the fact that the film of water between them serves as an effective buffer to protect such tiny particles against abrasion.

Quartz so commonly forms sand that, unless otherwise stated, quartz sand is generally understood. Many other minerals occur in sands, however, and the beaches of coral islands are in places formed of "coral sand" made wholly of calcium carbonate.

Silt and *clay* include the finest part of the land waste. This finely divided material coheres when wet, and when dry will crumble into dust. The distinction between silt and clay is largely one of fineness of subdivision, clay consisting of microscopic particles (less than 1/256 mm. in diameter) and *silt* of somewhat coarser sediment. Due to its most finely divided (colloidal) constituents, clay is plastic when wet whereas silt tends to crumble as would very fine sand. The common term *mud* is applied to either silts or clays and is used particularly where the separation of silt and clay is very imperfect.

The materials carried in solution form sediments wherever evaporation or chemical reactions cause their precipitation, or where organisms extract the dissolved minerals to form shells.

Volcanic ash, wind-blown dust, glacial till, and organic materials such as peat are other deposited materials that enter into the formation of special types of sedimentary rocks, but although of interest and local importance, they are not of such geological significance as those mentioned above.

Places of Deposit

In Chapter IV it was explained how streams transport their burden of sediments. The spreading of marine sediment by the waves and currents has been discussed in Chapter VIII. The conditions under which the sediments finally come to rest must now claim our attention.

Three major realms of deposition are generally recognized. The first is that of the land areas. The deposits formed there are known as *continental sediments* since they belong essentially to the continents, whether formed actually on the dry land or under the surface of its streams and lakes. The second realm is that of the intertidal zone and its deposits are known as *littoral sediments.* The third is that of the marine realm and its deposits, formed anywhere over the sea or ocean floors, constitute *marine sediments.* The distinction between these three realms is of such geologic significance that each of them deserves more detailed consideration.

Regions of Continental Deposits. — Although vast areas of the land surface are undergoing active erosion, about 10,000,000 square miles of the continents are now buried by accumulating sediments or by sedimentary deposits formed in the recent geologic past. These regions of deposition may be classified into four types in which the conditions of accumulation are strikingly different. These are (1) piedmont plains, (2) arid basins, (3) humid basins, and (4) great deltas.

Piedmont alluvial deposits are formed where streams debouch from relatively young and lofty mountain ranges. Here the velocity of the streams is rapidly checked as they enter the flatter piedmont belt and they drop a part or all of their load. During times of flood, when on leaving the mountains they are heavily laden with sediment, the streams spread widely beyond their channels and cover the adjacent country with layers of sand and clay. In so doing, they overflow into the lowest places, filling them with sediment, and by the continued action of this process through a long period of time they build up deposits of great thickness in the form of alluvial plains in front of the mountains. Striking examples are to be seen in the deposits that underlie the High Plains east of the Rocky Mountains, the Pampas east of the Andes, and the Indo-Gangetic Plain south of the Himalayas in India.

Since they are formed on the flanks of the mountains and well above sea level, these sediments will ultimately be destroyed as the erosion cycle progresses, unless preserved by deep downwarping of the piedmont region. Even now, for example, the very streams that formed the High Plains are stripping these deposits away again.

Desert deposits occupy great areas in the arid basins. Almost one-tenth of all the land surface of the world is embraced in interior basins, whose drainage is centripetal, with no outlet to the sea. Here the waste of the slopes constantly tends to move toward the deeper parts of the basins and accumulate in stratified deposits. At rare times of heavy rainfall the temporary streams spread widely over the lower slopes, shifting sand and gravel toward the center of the basins, while the finer material is swept on into temporary lakes (playa lakes) that form in the deeper depressions. Permanent lakes like Great Salt Lake or the Caspian Sea tend to be filled by the sediments of the larger and more permanent streams. Moreover, in times of dryness, sand is shifted by the wind. Except for wind-blown dust none of the sediment can escape and thus, through the continued action of rain wash and streams, the desert basins tend to fill with deposits. If sinking of the basins occurs while they are being silted up, the ultimate thickness of the sedimentary deposits thus formed may attain several thousands of feet. Deposits of arid basins commonly contain layers of salt and gypsum as well as thick beds of cleanly sorted dune sands. They are generally poor in fossils and exhibit light colors such as dun, buff, or pink. Some of them are red.

Deposits in humid basins also may be of vast extent. Where structural basins form rapidly in regions of abundant rainfall, great lakes are produced; but if the sinking is slow, filling may keep pace with subsidence and the region remain a swampy lowland. The second alternative is favored where the basin has bordering highlands undergoing rapid erosion, for then the incoming streams are heavily laden with sediment. A good example of this sort exists in the upper basin of the Paraguay River in South America where an area about 400 miles long and as much as 150 miles wide remains a labyrinth of lakes and swamps and channels in a low grassy plain. During the annual rainy season the whole area is flooded, but during the rest of the year only about one-fourth of it is covered with water. The river enters the lowland heavily burdened with sediments brought by tributaries from the Andes to the west and from the Brazilian highland to the north. The water leaves the basin fairly clear, having spread most of its sediments over the lowland to be mingled with organic matter from the decay of vegetation that flourishes abundantly in such places. This swampy condition, with resulting accumulation of river sediment, may be indefinitely maintained if the basin is a region of continued subsidence. Though deposited in water, such sediments are to be regarded as continental in origin, since they occur in hollows of the land surfaces. Such deposits have accumulated extensively in times past and seem to represent approximately the conditions under which the older coal-bearing rocks have been formed. The abundant organic matter gives such sediments dark colors.

Deposits formed in this manner are likely to escape destruction by later erosion because they lie only slightly above sea level.

A *delta* is a deposit of sediment built out by a stream into standing water. An idealized vertical section of a delta (Fig. 142) shows that it is partly above and partly below water level. Even the landward portion is extensively covered by fresh water when the stream is in flood and carrying its greatest load of sediment. At such times sand and silt are spread far and wide by the flood waters, forming nearly horizontal layers over the landward top of the delta. The sediment that reaches the shore is shifted by the waves and currents toward the sea, building out the shallow submarine surface like a vast embankment with a nearly horizontal surface near shore, giving way to a steeper slope beyond the line of most rapid building. The sediments that come to rest on the nearly horizontal upper surface of the delta, whether the landward or the subaqueous portion, are known as the *topset beds.* The material carried out over the inclined front of the delta constitutes the *foreset beds,* and that which is spread still farther, parallel to the sea bottom, forms the *bottomset beds.* The topset beds are partly continental but the foreset and bottomset beds are marine.

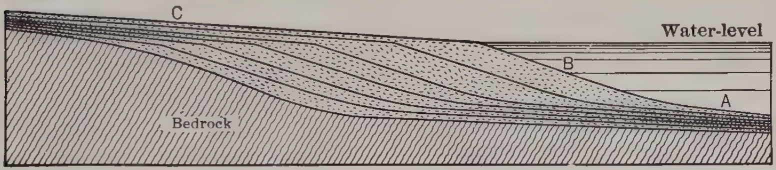


Fig. 142. — Section through a delta built into quiet water of constant level. *A*, bottomset beds; *B*, foreset beds; *C*, topset beds. (After Barrell.)

Regions of Littoral Deposits. — The total area of the present littoral zone in all the continents is probably not much over 60,000 square miles. On ordinary shores it is usually far less than 2 miles wide, though in great deltas like that of the Mississippi the salt-water marsh may be 25 miles or more in width and littoral sediments may accumulate rapidly thereon. The most remarkable extension of the existing littoral zone occurs in the so-called Runn of Cutch, a low flat on the southeast side of the Indus River delta that has an area of about 6000 square miles. It is so flat and so nearly at sea level as to be flooded with marine water during one season of the year when monsoon winds blow inland, and laid bare or covered by overflow from the Indus when the winds blow offshore. Some of the older mud-cracked marine formations may have formed under comparable conditions. In general, however, littoral deposits cannot be of great thickness, for, if the land is building out into the sea, they must give place to land deposits and be buried under them; or if the sea is encroaching on the land, they must yield to marine sediments and be covered by them. In either case, they stand a rather poor chance of permanent preservation, since upon slight uplift they are exposed to stream erosion and with stationary sea level the waves gradually encroach upon the shore zone and remove them. They are likely to be preserved only in subsiding large deltas.

Regions of Marine Deposits. — The greater part of the land waste finally comes to rest on the sea floor. Most of this terrigenous (land-derived) sediment accumulates in the neritic zone but the finest muds can be transported even beyond the edge of the continental shelf and are found as much as 200 miles from land. Coarse materials accumulate in shallow water near the shore whence they are supplied, for there alone are the currents strong enough to distribute gravel and coarse sand. Fine sands extend in places to a depth of 200 or 300 feet or even more and beyond them the terrigenous muds spread as a continuous blanket not only to the continental margin but part way at least over the continental slope, where they grade imperceptibly into the oozes of the ocean floor. It is improbable, however, that much land-derived mud reaches the abyssal region. While in general there is a seaward gradation from coarse to fine sediment, it must not be supposed that all the muds are deposited in deep water, nor that there is always a shoreward phase of sandy sediment between the land and the place where mud deposition begins. In many protected embayments, as the Baltic Sea or Chesapeake Bay, muds, even of the finest sort, are now being deposited right up to the shore line.

Calcareous deposits of the neritic zone form chiefly in shallow water where clastic (fragmental) sediments are scarce or absent and where lime-secreting organisms grow luxuriantly. At the present time limy deposits are forming chiefly about the tropical coral islands and over warm shallow submarine banks such as that bordering Florida on the south and the Bahama Islands on the west. It has sometimes been wrongly inferred that marine sediments grade in depth from gravel or sand through muds to limy ooze, and it should be emphasized, therefore, that the limestones now exposed on the continents represent deposits in shallow, not deep water.

The vast deposits of red clay and organic oozes that cover the deep ocean floor have been described in the previous chapter. They are of enormous extent, but since such deposits are essentially lost to the continents and only rarely have been elevated into lands their geologic interest is limited.

Consolidation of Sediments

Stratified rocks are merely sediments consolidated into stone. Commonly the appearance of the rock clearly suggests this origin, but some of the older rocks have been modified so greatly that the resemblance to sediments would be obscure if it were not for their stratification and entombed fossils. The consolidation is the result of several factors, chief of which are *pressure* and *cementation.* As the sediments accumulate, the upper layers bear upon all those buried below. If the specific gravity of loose sediment is 2.5, the pressure will increase by fully 1 pound per square inch for each foot of depth. At a depth of 1000 feet, therefore, the weight of the overlying deposits is at least 1000 pounds per square inch. As a result of such compression, sediment is made more compact as the particles are squeezed tightly together and most of the water is pressed out.

Cementation results from the deposition of mineral matter in the spaces between the grains of sediment. The most common cementing substances are calcium carbonate, silica, and iron oxide. These cements may be introduced in solution in the water that fills the pore space when the sediments are deposited, or they may be introduced later by percolating ground water. Some sandstones have but little cement and therefore crumble readily into loose sand when exposed to the weather, but others have the voids almost completely filled with cement (Fig. 143) and have been thus converted into firm rock.

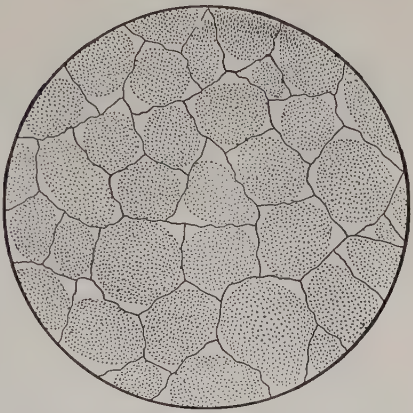


Fig. 143. — Microscopic section of a firmly cemented sandstone. The dotted areas are the rounded sand grains and the clear areas represent silica deposited between the grains, binding them into rock.

The interior heat of the Earth rising into such masses of sediments may aid in some degree to consolidate them by quickening the chemical activity of the diffused waters which deposit the cement. And finally, since the conversion of sediments into rock must be a slow process, *time* is an important element. Thus in general the more recent sediments, where they have been exposed, are softer and more friable than the older ones.

Kinds of Sedimentary Rocks

The different kinds of sedimentary rocks depend mainly upon the nature of the sediments from which they are formed. The chief types of sediment and the equivalent sedimentary rocks are as follows:

Sediments	Compacted Strata, as Rocks
Gravel	Conglomerate
Sand	Sandstone
Silt and clay (mud)	Shale
Calcareous mud and sand	Limestone

It must not be imagined that the different kinds of rocks mentioned above are always sharply defined from one another as wholly distinct types. Just as muds grade through sand into gravel, and pure calcareous deposits into silts or clays, so the various rocks formed from them grade into one another. In the description of these impure or mixed types of rocks, those that were muddy or clayey sediments are termed *argillaceous* (from the Greek *argillos*, clay). Similarly, sandy ones are described as *arenaceous* (from the Latin *arena*, sand) and limy sediments or rocks are described as *calcareous* (from the Latin *calx*, lime). These three adjectives are used in the discussion either of sediments or of the rocks derived from them.

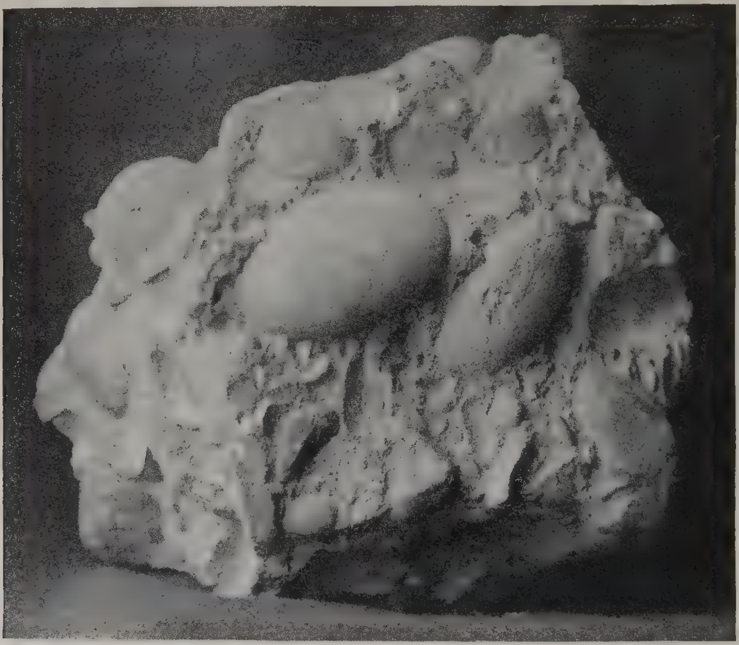


Fig. 144. — A piece of conglomerate, shown about half natural size.

The chief types of sedimentary rocks deserve more detailed discussion.

Conglomerate. — A typical conglomerate is shown in Fig. 144. It consists of rounded pebbles set in a matrix of well-cemented sand. Generally the pebbles and boulders in conglomerates are made of quartz, but they may consist of fragments of any kind of rock. The fragments vary in size from a small fraction of an inch to many feet in diameter. They may be packed together with little fine matrix or mingled with any proportion of fine material. If the pebbles are few and scattered through a sandy matrix, however, the rock should be called a *pebbled sandstone* rather than a conglomerate.

The pebbles and boulders may be well rounded or, if they have suffered but little transportation, they may be more or less angular. If they are distinctly angular the rock is a variety of conglomerate known as *breccia.*

Arkose is a special variety of conglomerate or sandstone containing much feldspar. Its occurrence indicates that the component material was not long exposed to weathering before it was deposited, and therefore probably was not transported great distances. Arkose is more commonly of continental than marine origin, and its formation is favored by the breaking down of granitic rocks in cold or arid climates where rock decay is inhibited.

Sandstone. — Sandstones are composed of cemented sand grains, which consist most commonly of quartz. Many sandstones are quite even in grain, but there are hybrid varieties grading on the one hand into conglomerate, and on the other into shales. In red and brown sandstones the cement is mainly oxide of iron but in white, buff, or gray varieties it is usually either silica or calcium carbonate. Sandstones generally are porous and the interspaces amount to as much as 30 per cent of the total volume. For this reason they are favorable reservoirs for artesian water and petroleum.

Shale. — The indurated equivalent of silts and clays is shale. Most shales possess a more or less thinly laminated structure resulting from closely spaced, parallel bedding planes. Such rocks tend to split easily along these natural planes of stratification. If the laminae are thin and flexible the shale is said to be fissile. On the other hand, some shales occur in layers of considerable thickness that show no subdivision by bedding planes and tend to break into blocks instead of laminae. Such massive or blocky shales are sometimes designated *mudstones.* Each thick layer represents rapid deposition at the time of a single storm or flood.

Shales are soft and generally weak rocks, crumbling readily into small chips. They show a great variety of colors as do the muds from which they form. Unlike sandstone, they tend to be impermeable to water. Since they are composed to a greater or less degree of clay they yield a strong clay odor. Shales occur in both continental and marine deposits.

Black shales constitute a distinctive type of sedimentary rock commonly associated with coal beds and also occurring abundantly in marine deposits, where they are important as source rocks in the geological formation of petroleum. The dark color of these shales is due to carbon derived from the incomplete destruction of organic matter buried with the mud. An abundant supply of organic matter and stagnation of the water appear to be the essential requirements for black shale formation, the stagnation leading to a deficient supply of dissolved oxygen so that decay is inhibited. Depth is not a vital factor, though obviously any sharply depressed areas or hollows in the sea floor are likely to be passed over by bottom currents and to become stagnant black mud holes. Black muds are now accumulating in both shallow and deep water. Where black muds are forming, hydrogen sulphide (H_2S) is usually generated and it in turn reacts with iron salts in the sea water to form pyrite (FeS_2), which is precipitated in the form of concretions in the black shale, or very commonly in the form of replacements of fossil structures.

Limestone. — Limestone is the consolidated equivalent of limy ooze, calcareous sand, or shell fragments. It is composed of the mineral calcite or calcium carbonate ($CaCO_3$). Pure varieties of limestone are white or light gray, but impure varieties are buff, brown, red, dark gray, or even black. If fine-grained and nearly pure, the rock is compact and tough and forms one of the strongest of structural stones; but with an admixture of clay, limestones grade into calcareous shale which weathers readily.

Chalk is a soft, porous variety of limestone. Some chalks are composed largely of the microscopic shells of Foraminifera, the tiny animals that also make the globigerina ooze of the ocean floor. These minute shells are extremely fragile and commonly occur much broken. In addition to the shell fragments there is normally a matrix of fine particles of calcium carbonate that has been considered to be a chemical precipitate. Some chalky deposits, as those of Kansas and Alabama, are made up largely of such finely divided material, with an admixture of clay, and show but few of the foraminiferal shells.

Chalk was once supposed to represent deep-water deposits like the modern oceanic oozes; but in spite of the fact that globigerina ooze will probably form chalky deposits, it is now known that the great chalk beds of the present lands include shells of shallow water animals and are associated with coarse-grained sediments that could not have been washed into deep water.

Coquina is a variety of limestone made up of shells and coarse shell fragments heaped together and loosely cemented.

Dolomite is a rock formed largely of the magnesium-calcium carbonate mineral, dolomite, $CaMg(CO_3)_2$. It may be considered a special variety of limestone. In fact, there are probably all gradations between calcite and dolomite limestones, and the mixed types are more common than pure dolomite. Dolomite resembles compact calcite limestone in most of its characters and is commonly regarded as limestone by others than specialists. It is of great extent and importance in the older rocks, especially. The Dolomite Alps of the eastern Tyrol, for example, have been carved out of dolomite formations 3000 to 4000 feet thick, and equally great masses of this kind of rock occur in Alabama and in western Texas.

Limestone represents the calcium carbonate that is carried in solution to the sea and there precipitated; but the conditions for its deposition are less obvious than those under which the fragmental sediments are formed and there is still much to learn about the relative importance of the different agents that precipitate the limy sediments.

The soluble bicarbonate of calcium is very unstable and can easily be made to give up one molecule of its CO_2 according to the formula $H_2Ca(CO_3)_2 = CO_2 + H_2O + CaCO_3$; but when this happens the corresponding molecule of $CaCO_3$ becomes insoluble and is precipitated. Recent investigations have shown that the warm shallow marine water is essentially saturated with calcium carbonate whereas the colder, deeper water is undersaturated because of its richness in CO_2. Wherever the deeper water rises in ocean currents to flow over shallow submarine banks in tropical regions the rise in its temperature tends to drive off part of the CO_2 and to leave the warmed water supersaturated with $CaCO_3$. These places afford the most favorable conditions for animals and plants that are prodigal in their use of carbonates to form shells or skeletal structures. Corals thrive and form reefs on such tropical shoals and with the corals are associated lime-secreting algae and a host of shell-forming creatures.

Impressed by the limy mud and coral sand formed about coral reefs, the earlier naturalists attributed to corals a predominant rôle in the precipitation of the limy deposits. Many limestones, however, present little evidence of a coralline origin and recent investigations have shown that modern calcareous sediments of great extent are being precipitated directly as a fine ooze or mud. One of the most extensive areas of modern limy sediment is that of the Bahama Banks south of the Florida

Straits. The total area of this great shoal is over 7000 square miles and the average depth of its water is less than 20 feet. The bank is formed of limestone and great areas of its surface are mantled by fine white limy ooze.

Minor Sedimentary Rocks. — *Coal, iron ore, rock salt, gypsum,* and *chert* or *flint* are sedimentary rocks of special interest, but from the geologic point of view they occur in volumes so limited in comparison with the enormous bulk of the shales, sandstones, and limestones that they are of little importance considered merely as rock masses.

Characteristic Features of Sedimentary Rocks

In addition to the stratification which these rocks always exhibit, other features are commonly displayed which throw light upon the geologic history of the rocks. Among these features are fossils, mud cracks, ripple marks, and cross-bedding.

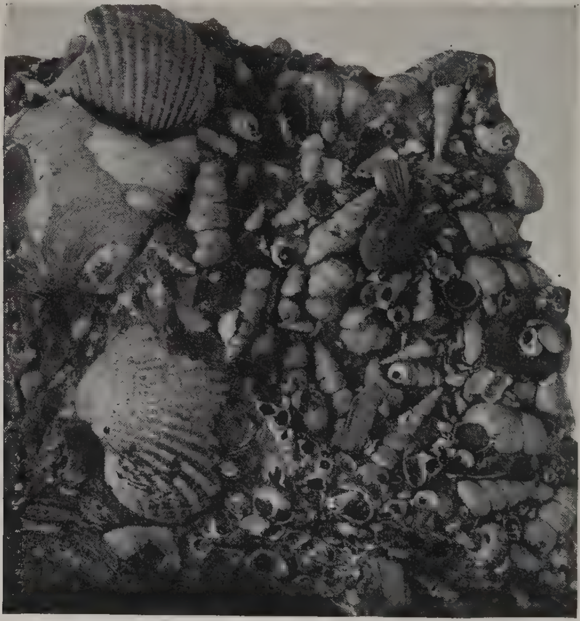


Fig. 145. — Fossil shells in stone.

Fossils. — Fossils (Latin *fossiles*, from *fodere*, to dig up) are the remains or imprints of animals or plants that were buried with the accumulating sediments (Fig. 145). The organisms are rarely preserved entire; generally only the hard structures, such as shells and bones, endure. These parts may be preserved without change, but commonly the original material is replaced, one molecule at a time, by mineral matter so that the fossil becomes a solid stony object or *petrifaction.* Not uncommonly the organic structure is wholly removed, leaving only an imprint or hollow mold in the rock. *Natural molds* of this sort are not infrequently filled by subsequent mineral deposits which then form *natural casts* of the organic objects. Finally, footprints or trails made by animals crossing soft mud are preserved under favorable conditions.

Since most fossils represent the animals and plants that were living where the sediments accumulated, they throw much light on the conditions that prevailed during the deposition of the fossiliferous rocks. They show, for example, whether the beds were laid down on the land or beneath the sea, since continental sediments contain the remains of land plants, bones of land animals, or shells of river clams, whereas marine sediments contain representative marine shells. Something of the climate of the region at the time of deposition may be indicated also; for example, fossils of tropical plants and animals, such as palms and alligators, are found associated in certain formations of the " Badlands " of South Dakota. The manner in which fossils are used in deciphering the past history of the Earth and its inhabitants is considered in detail in the second part of this book.

Mud Cracks. — Soft muddy sediments left exposed after the recession of flood waters shrink and crack into characteristic polygonal blocks like those shown in Fig. 146. These desiccation fractures are known as *mud cracks.* Further exposure to air and sun bakes and hardens the blocks of mud. During the dry season wind-blown sand or silt may cover the surface, filling the cracks with sediment coarser than the mud-cracked layer. In this way the form of the polygonal blocks is preserved even after the return of later flood waters. After the whole series of deposits has subsided and become hardened into rock, the layers of shale and sandstone may later be exposed, exhibiting these " fossil " mud cracks on the bedding planes. Obviously, the conditions that favor the formation and preservation of mud cracks are also ideal for the preservation of footprints of animals that have crossed the soft mud soon after its deposition, and therefore mud cracks and footprints are frequently associated.

The most favorable places for the occurrence of mud cracks are on the flood plains of large rivers, the landward portions of great deltas, and the wide flat shores of shallow interior lakes that shrink or disappear during the dry season. In spite of its alternate wetting and drying, the littoral zone is unfavorable for the formation of mud cracks because the mud has not sufficient time to dry out thoroughly before the return of the tide. Mud cracks may, however, form to a limited extent at the upper margins of estuaries where the spring tides reach only for a few days in each month. The mud cracks can form only when the sediment is exposed to the air and allowed to dry and shrink. They are generally lacking therefore in marine deposits.

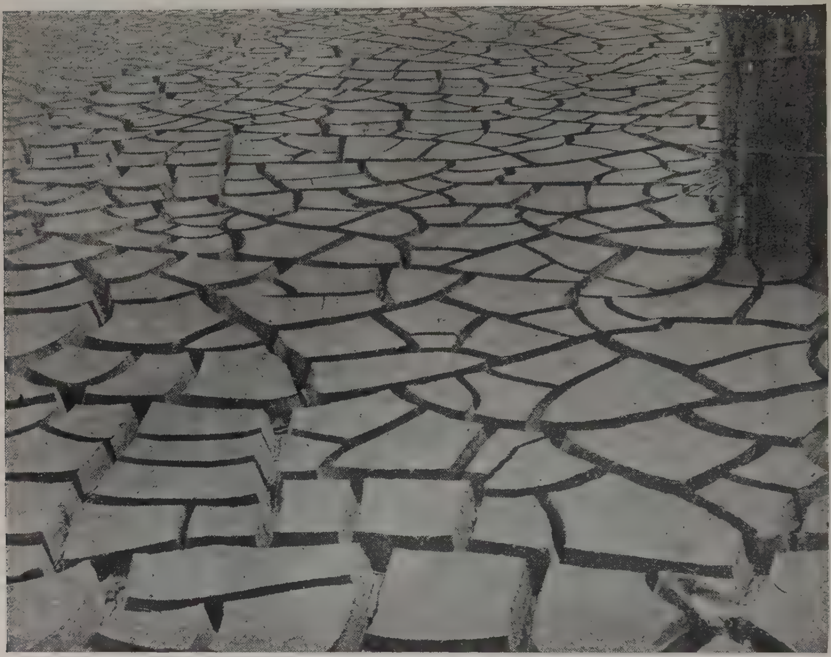


Fig. 146. — Modern mud cracks formed on the delta of the Colorado River. (U. S. Geol. Surv.)

In the early Paleozoic formations, however, there are well-known examples of mud-cracked marine limestones. They represent very special conditions that may find an analogy at present in the Runn of Cutch or in the Bahama Islands. During the hurricanes of the spring of 1928 the lower islands and the borders of the larger islands of the Bahama group were flooded by the seas that swept over the Bahama Banks (see p. 193). It was later observed that many square miles of this sediment were mud-cracked. In places limy sand and fragments of marine shells have since blown over it. It is quite possible that in the past similar conditions have permitted the covering of extensive low coastal lands by marine sediments laid down during exceptional hurricanes beyond the normal confines of the sea.

Ripple Marks. — Where currents sweep granular sediments along the bottom, the surface of the deposit develops parallel ridges resembling the ripples on the surface of a pool of water. These are known as *ripple marks.* They are also formed on the land where sand is shifted by the wind, as on sand dunes, or under running water. Ripples thus produced retain their form and migrate slowly with the current, because the sand grains are rolled up the upstream or stoss side and fall down the leeward slope. Current-formed ripples, therefore, have an unsymmetrical form, the stoss side being a gentle slope and the opposite side steeper. The direction of flow of the current is thus autographed in the form of the ripple marks (Fig. 147).

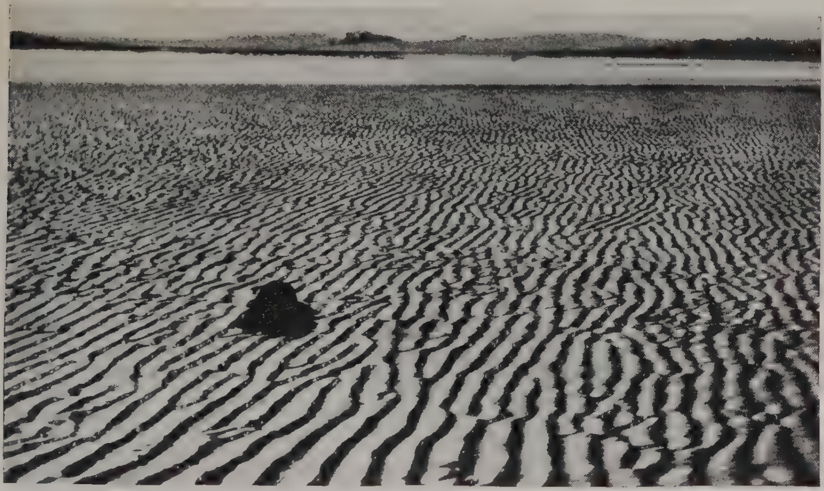


Fig. 147. — Current-formed ripple marks exposed at low tide near Windsor, Nova Scotia. The movement of the current was from left to right. (Geol. Surv. of Canada.)

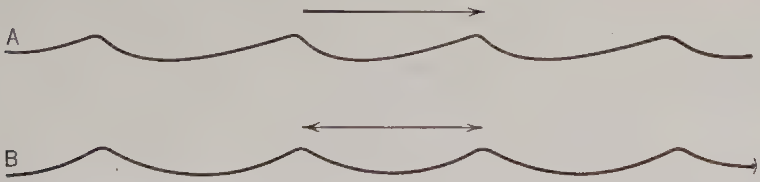


Fig. 148. — Unsymmetrical profile of current ripples (*A*) contrasted with the symmetrical profile of oscillation ripples (*B*).

Where oscillatory waves touch bottom they also develop ripples as a result of the to-and-fro motion of the bottom particles; but oscillatory ripples, in contrast to current-formed ripples, are symmetrical (Fig. 148).

Current ripples may be formed wherever currents disturb sandy or silty surfaces, whether it be wind currents on the land, the currents of running streams, or any of the currents in the seas. Waves of oscillation, on the other hand, occur only under standing water and in depths touched by wave action. Ordinary storm waves in the sea are ineffective below 200 or 300 feet, but exceptionally the bottom is rippled to depths of 600 feet or more.

Cross-bedding. — In many deposits of coarse detritus, such as conglomerate and sandstone, the layers of particular beds are inclined to the general planes of stratification at considerable angles (Fig. 149). This structure is known as *cross-bedding.* It is produced where sand is shifted by either wind or water currents in such a way as to be spilled down the front of an advancing deposit; as the foreset edge of a delta, the front of a gravel bar in a stream, the front of a sand dune, or merely the front of a current ripple. In any case the individual layers or laminae come to rest on a slope inclined to the general surface of deposition.

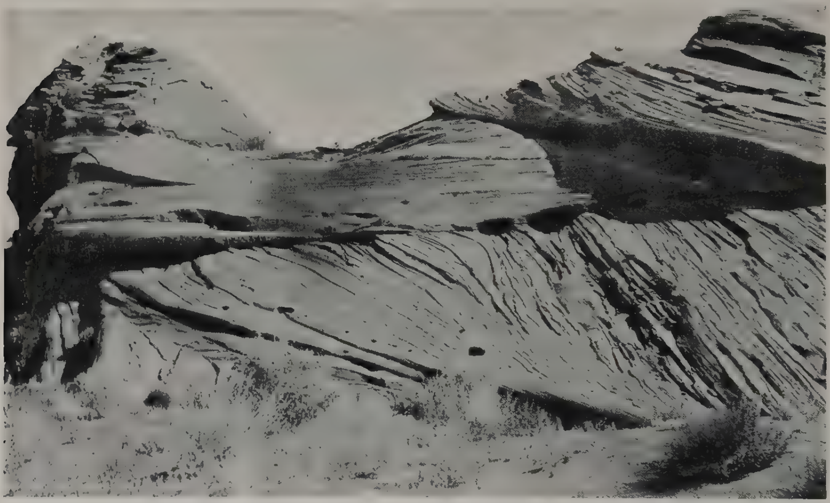


Fig. 149. — Cross-bedding of the type characteristic of sand dunes. De Chelly sandstone, Arizona. (U. S. Geol. Surv.)

The scale of the cross-bedding may be great or small. In dune sands single cross-bedded layers are commonly tens of feet thick but in silts shifted by small ripples the cross-bedded layers are fractions of an inch thick. Since the inclination or foresetting of the laminae is always downstream, it is possible to infer the direction of the currents that moved the cross-bedded deposits.

The cross-bedding produced in dune sands is one of the most distinctive types. The laminae are inclined first in one direction and then in another as a result of the repeated change of direction of the winds. Moreover, the front slope of the dune is not straight but curved where the sand rolls forward at its base as a result of eddies in the wind on the leeward side of the dune. As a result, the cross laminae, developed on a large scale, descend in curves that become tangential to the bedding planes below.

Concretions. — Stratified rocks in many places contain inclusions called concretions. These objects differ in composition from the enclosing rock and are generally rounded or nodular in form; some are quite spherical, others flattened, ovate, elongated, ring-shaped, or compound; still others exhibit odd and fantastic shapes. They range in size from a fraction of an inch to many feet in diameter. Ordinarily they are formed from one of the minor constituents of the rock; thus, in chalk and limestone they are composed of silica; in sandstone, of iron oxide or carbonate of lime; in shale, of calcium carbonate or sulphide of iron. Although some are pure they commonly contain large amounts of the inclosing rock material, and the planes of stratification of the rock can be seen passing through some concretions.

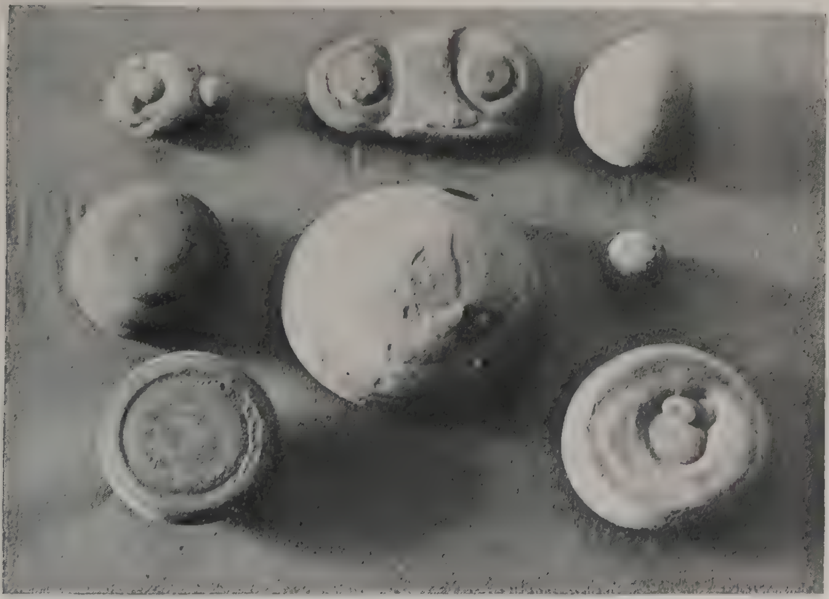


Fig. 150. — Concretions from clay beds, Long Island.

Their origin appears to lie in the solution of some of the minor constituents in the rock and the redeposition of this material around certain centers as nuclei. Very commonly a concretion contains at its center a fossil and in such cases it appears that the products of decay of the buried organism have caused the precipitation of the mineral matter. Remarkable imprints of fern leaves, insects, and marine animals are obtained by splitting such concretions. The shells and bones even of large animals are found in some concretions. Others appear to have formed about nuclei of inorganic substance, such as grains of sand, and still others show no definite nucleus of any sort. Iron oxide concretions not uncommonly are hollow or have only a partial filling of loose sand.

Colors of Sedimentary Rocks

Sedimentary rocks present a great variety of colors, some of which help to indicate the conditions under which the sediments accumulated. Sandstones usually are light-gray, greenish-gray, buff, brown, or red; shales generally light-gray, dark-gray, greenish-blue, red, purple, maroon, or black; limestones are nearly white if pure but gray, buff, brown, pink to red, or black if impure. Some of the more recent deposits of clay are nearly white or have delicate shades of pink or lavender. The coloring matter is usually carbon or oxides of iron.

Black or dark-gray sediments owe their color essentially to the black carbon resulting from the partial destruction of organic matter that was buried with the sediment. The black shales are especially rich in organic matter, and coal represents concentrated deposits of this nature.

Oxides of iron are the great coloring agents in nature. Iron is the fourth most abundant element in the Earth's crust and iron-bearing minerals are disseminated throughout almost all rock masses. There are few sediments, therefore, that are completely wanting in iron. In the igneous rocks the iron occurs generally in the ferrous state combined in the silicate minerals and in this condition is not strongly colored. In weathering, however, the iron becomes oxidized and then takes on bright hues of yellow, brown, or red like those displayed by rusting iron. Iron oxide has the capacity of combining with a variable proportion of water to form hydrated iron oxides and the shade of color from yellow through brown to red seems to depend largely on the proportion of adsorbed water. The dehydrated ferric oxide (Fe_2O_3) is of deep red color.

Stratigraphic Relations

Much light is thrown on the geologic conditions under which a group of strata was formed by a study of the distribution of the beds, their relation to other rock masses, and, especially, the distribution of the fine and coarse sediments in the group. Some of the more important of these stratigraphic relations are discussed in the following paragraphs.

Relative Age of Beds. — In view of the origin of stratified rocks as superposed layers of sediment, it is evident that each layer is younger than the next below. Except where the rocks have been overturned, or broken and thrust out of their normal sequence as in some mountain zones, the youngest is at the top and the oldest at the bottom of any group of strata. Many conclusions as to the historical sequence of geological events rest upon this evident and fundamental law of stratigraphy.

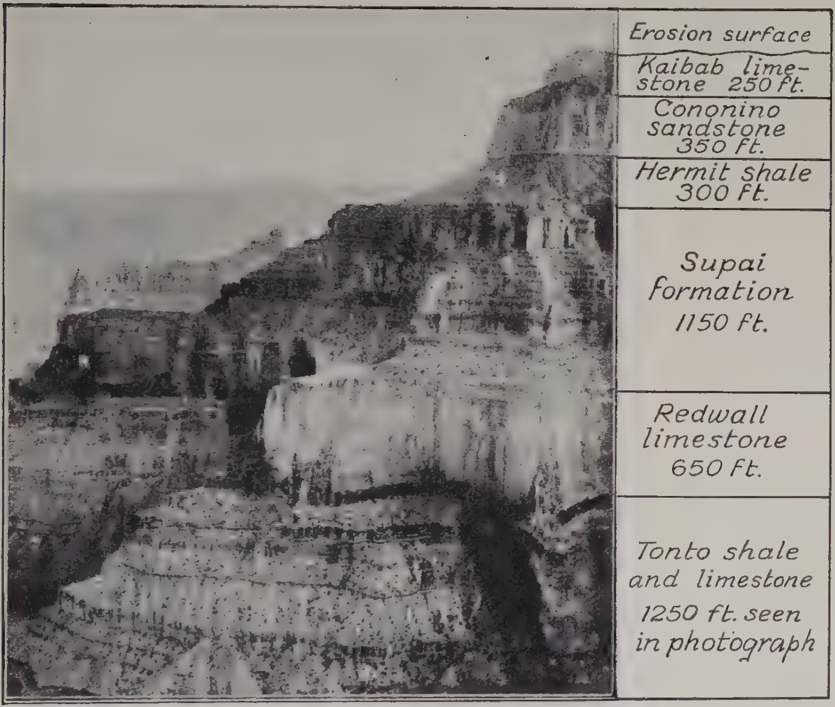

Erosion surface
Kaibab limestone 250 ft.
Cononino sandstone 350 ft.
Hermit shale 300 ft.
Supai formation 1150 ft.
Redwall limestone 650 ft.
Tonto shale and limestone 1250 ft. seen in photograph

Fig. 151. — Part of the south wall of the Grand Canyon of the Colorado River near El Tovar, showing six great geological formations, each of which is marked with the initial letter of the name shown in the explanatory diagram. (U. S. Geol. Surv., and Ariz. Bureau of Mines.)

Grouping of Strata into Formations. — A group of similar strata, closely related in their development, constitutes a geological *formation.* The view of the Grand Canyon wall seen in Fig. 151, for example, shows six great formations. The lowest includes intergrading beds of sandy shale and shaly sandstone more than 1000 feet thick; the next is a rather pure and homogeneous limestone that outcrops in vertical cliffs along the middle of the Canyon wall; the third is a thick group of shaly and sandy beds; the fourth a homogeneous shale; the fifth a pure, cliff-making sandstone; and the uppermost a limestone. During the time of the formation of the lowest group of strata, conditions of deposition must have remained nearly uniform and the processes supplying fine sand and mud to this region presumably operated with more or less continuity. A different set of conditions then came into existence and endured while the thick limy sediments of the second formation were produced. Eventually conditions changed again and yet again to give rise to the higher groups of strata. Thus each of the formations is the product of a distinct set of formative processes.

It is the universal custom to give rock formations geographical names. Thus, certain limestone beds well exposed about St. Louis are known as the *St. Louis formation* and the name is extended to them wherever they can be traced or identified. Likewise a group of beds forming river bluffs about Kansas City and extending far away north, east, and south, constitutes the *Kansas City formation.*

Intergradation of Different Kinds of Sedimentary Rocks. — Sediments vary from place to place in their coarseness and in their composition. Gravel and coarse sand accumulate where transporting currents are able to carry them, but grade laterally into finer sand where the currents slacken. Fine sand in turn grades into mud in the sheltered places and in quiet depths of water. The limy deposits of clear shallow sea bottoms grade eventually into muddy or sandy deposits near degrading lands. The change from one type of sediment to the next is always gradational and seldom abrupt. It is inevitable, therefore, that sedimentary rocks of one kind grade into contemporaneous strata of different composition. In general, coarse sediments grade most rapidly into other types of deposit and are most localized and irregular in their distribution. Thus conglomerates, which mark old stream channels or shore zones, generally form irregular linear deposits extended parallel to the stream course or the shoreline but in other directions grading into sandstone within a few miles at the most. Coarse sandstones generally show a similar inconstancy, but single beds of shale or limestone may cover thousands of square miles.

In a basin receiving sediments for a considerable time, gradations will occur not only in a lateral sense, but from bottom to top of the deposit, as the erosion cycle progresses. If, for example, a land area were uplifted and then slowly degraded the sediments derived from it and deposited in adjacent depressions would show a regular cycle of deposition (Fig. 152), with coarse material near the base succeeded by finer and finer sediment. The sedimentary cycle is a corollary of the erosion cycle. In the youthful and submature stages of dissection the streams can transport abundant coarse sediment but, as the land is worn lower, the streams flow more slowly and carry only fine material, and finally, late in the cycle of erosion when mechanical sediments are reduced to a minimum, calcareous deposits may form near the shore. If the land area is then rejuvenated, rapid erosion begins again and a new cycle of deposition is inaugurated, clastic sediments coming to rest upon the limestone.

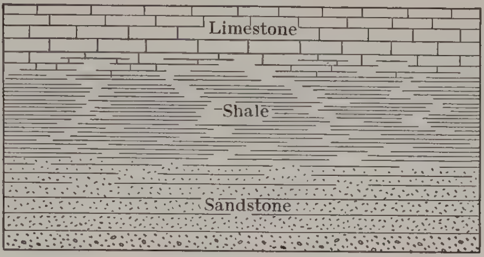


Fig. 152. — To indicate the normal succession of sedimentary deposits derived from a degrading land mass.

Actually, the sedimentary cycle is seldom if ever brought to completion in the simplicity pictured above, for it may be interrupted by many factors as, for example, uplift of the land, increased erosion due to climatic changes, uplift of the region of deposition, or change in the course of marine currents.

Overlap. — Where the sea (or the area of deposition) gradually encroaches upon a land surface, the beds have the relations shown in Fig. 153, each layer overlapping the next below and extending some distance beyond it so as to thin out against the old land surface. This relation, known as *overlap*, is of common occurrence and great geologic importance. It should be noted that the sands, being near-shore deposits, grade laterally into muds. As exposed in the field, the true bedding might be indistinct and the entire sloping deposit of sand above the old land surface might be wrongly interpreted as a unit layer formed at one time

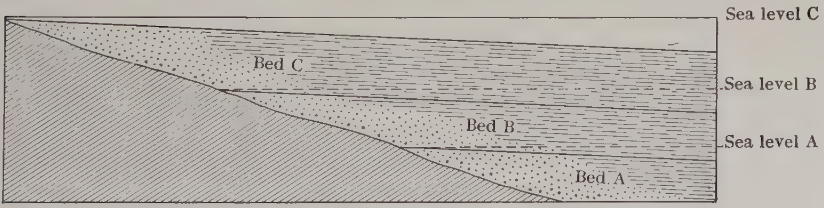


Fig. 153. — To illustrate the progressive *overlap* of beds *A*, *B*, and *C* as a result of the rise of sea level by three corresponding stages. Note that the shoreline has moved progressively to the left. The normal seaward inclination of the beds is exaggerated. Conventions for sandstone and shale same as in Fig. 152.

Extent and Form of Sedimentary Formations. — In view of the discontinuous nature of the regions of deposition it is evident that no geologic formation can be world-wide in its extent. Beds of mechanical sediment, as sandstone and shale, imply land surfaces from which they were derived, and basins in which they were laid down; obviously their areal extent must be limited by the borders of the basins next to which the sediments thin and disappear. In geometrical form a group of sediments is relatively broad and sheet-like, consisting of subparallel layers or strata. If deposited in a circumscribed basin the deposits are roughly lenticular and thickest in the deepest (or most rapidly subsiding) part of the basin; but if laid down along an open sea coast the form of the deposits tends to be wedge-like, thickest near shore and progressively thinner toward the ocean.

Unconformity. — One of the most important stratigraphic relations is that of unconformity, a subject that is treated in Chapter XII.

CHAPTER X

VOLCANOES AND VOLCANISM

Various agencies that modify the surface of the Earth, such as the atmosphere, the water in its several forms as rivers, seas, and ice, derive the energy that enables them to do their work from a source exterior to the Earth: from the sun. For without the sun these activities would cease and the Earth's surface would be inert. Toward these agencies the Earth is passive, except as it adds the force of gravity to help them in their work.

We have now to consider a set of agencies that also are modifying the Earth's surface, but whose energy is derived from sources within the Earth itself. So far as we can judge they are due either directly to the interior heat of the Earth or to changes going on within the Earth that produce heat. We shall describe first the results that they accomplish at the surface, and then inquire into their origin.

VOLCANOES

General Description. — A volcano is a hill or mountain composed of materials amassed around a vent through which they have been ejected from the Earth's interior in a highly heated or molten condition. The typical volcano is conceived of as a steep conical mountain with a pit-like crater at its top, from which issue from time to time gases, ashes, bombs, and flows of molten rock called *lava*. The ejection of material is termed an *eruption*, and to the human mind volcanic eruptions are perhaps the most impressive of all geological phenomena, from the immensity of the forces displayed, the magnitude of the results achieved, and the disastrous consequences that they frequently entail. Volcanoes vary widely from the typical form: some are low and of gentle slopes, others high and steep; some are elongated and irregular in shape; and some craters are at the side instead of the top, and they are variable in shape.

In size volcanoes range from small cones 100 or 200 feet high to some of the loftiest mountains on the globe. Thus certain of the highest peaks of the Andes are volcanoes; some of these are still active, as Cotopaxi in Ecuador, 19,600 feet high, with a crater half a mile in diameter and 1500 feet deep, whereas others, like Aconcagua (23,000

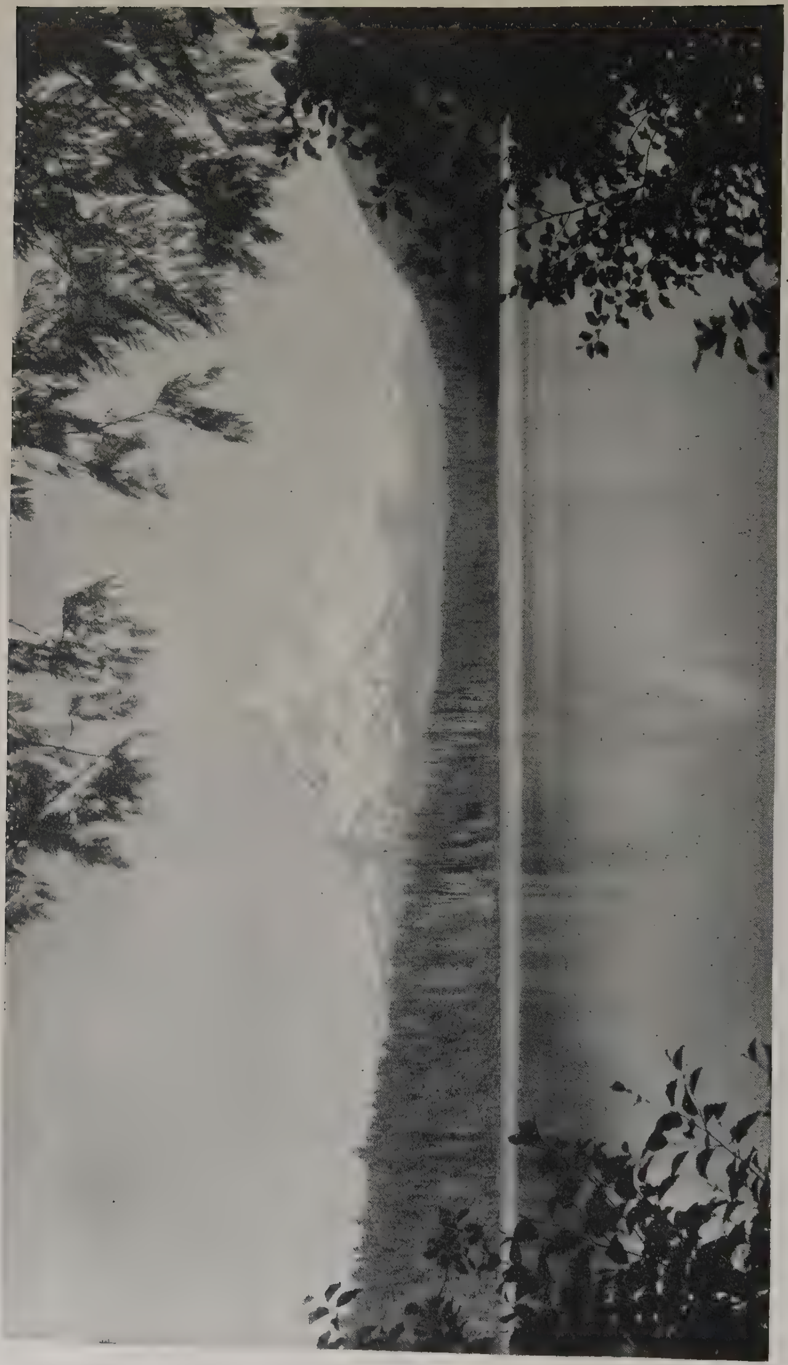


Fig. 154. — Mt. Hood, Oregon. A composite volcanic cone, somewhat modified by glacial erosion. (A. M. Prentiss.)

feet) and Tupungato (21,500 feet) on the border between Chile and Argentina, and Chimborazo (20,500), in Ecuador, which apparently have no craters and are not now active, have become extinct in the recent geologic past. These volcanoes are built upon a dissected platform of much older rocks, above which they rise 10,000 to 12,000 feet; but the volcanic masses of the Hawaiian Islands rise from the bottom of the Pacific Ocean at depths of 14,000 to 18,000 feet, and their highest summits project 14,000 feet above sea level, thus making the total height 30,000 feet. In the United States the higher peaks of the Cascade Range, beginning on the north with Mt. Baker (10,703 feet), Mt. Rainier (14,408), and Mt. Adams (12,307) in Washington; Mt. Hood (11,225) in Oregon (Fig. 154); and Mt. Shasta (14,162) in northern California, are volcanoes, which are now dormant or have recently become extinct. Lassen Peak (10,460 feet), standing at the south end of the Cascade Range, is the only active volcano within the continental United States. Mt. Etna, on the coast of Sicily, rises 11,000 feet above sea level, and the diameter of the base of its cone is 30 miles. Its lower slopes are gentle and studded with many small, or "parasitic," cones.

Character of Eruptions. — Three kinds of material are ejected from volcanic vents: gases, liquids consisting of molten rock, and solid material in the form of fragments; and the nature of a volcanic eruption depends largely on the proportions of these three things. If the eruption is violent and explosive, then the gases have been the chief factor in its production, and solid fragmental material is the main product; if, on the other hand, the eruption proceeds quietly, liquid rock, or lava, is the main product, and the gases play a less important rôle. We may roughly classify volcanic eruptions into those which are *explosive* and those which are *quiet.* When we classify actual volcanoes according to this difference in operation, we quickly find that, although there are good examples of both types, many, perhaps the majority, are *intermediate* in their character; that is, at times they erupt violently and at other times they quietly discharge flows of lava. In many volcanoes during a quiescent stage there appears to be a gradual accumulation of pressure, the lava rises in the conduit, and eventually the eruption begins explosively. Great quantities of gases mingled with dust and stones are ejected; and, the pressure being thus largely relieved, the explosive phase is succeeded by a quieter one in which the lava escapes through rents in the cone and flows out on its exterior.

Explosive Type. — The most extreme eruptions of this type take place as appallingly disastrous explosions. Enormous quantities of gas are suddenly ejected into the atmosphere, so thickly mingled with comminuted rock (dust and ashes), as to form vast outrushing and expanding clouds of dense appearance and dark color (Fig. 155). The greatest known explosion was that of Tamboro, on Sambawa Island east of Java, in 1815. Thirty or more cubic miles of material were blown into the air. The ashes produced darkness for three consecutive days for a distance of 300 miles from the volcano and dust fell over an area of 1,000,000 square miles. Krakatoa, a volcano in the Strait of Sunda near Java, exploded in August, 1883. After premonitory outrushes of gas for some time, the great explosions occurred, which blew into the air over a cubic mile of material in the form of dust and ashes. This rose as a vast dark cloud 17 miles into the atmosphere, by its denseness completely hiding the sun over an enormous area. The noise of the terrific detonations was heard as far as Australia; and the disturbance in the atmosphere was registered by barometers over the whole world. Huge waves, up to 100 feet above tide, were generated in the sea and rushed along the low-lying coasts of Java and Sumatra, sweeping far inland and destroying towns, villages, and the lives of nearly 40,000 people; these waves were perceptible 3000 to 4000 miles away.

In May, 1902, from the volcano of Mont Pelée on the island of Martinique in the West Indies, and almost simultaneously from that of Soufrière on St. Vincent 90 miles away, after small premonitory symptoms, violent explosive eruptions took place. No lava was poured out, but the intensely heated gases were so thoroughly charged with incandescent particles of rock that the heavy, fiery clouds rushed down the mountain slopes to the sea. Destroying all life in its course, the cloud on Martinique swept through the town of St. Pierre and immediately destroyed it, together with its 28,000 inhabitants. On St. Vincent 2000 people perished and a broad tract of country was devastated. For many months after, Mont Pelée continued to eject at irregular intervals these incandescent clouds (Fig. 156).

Intermediate Type. — Probably most volcanoes belong, or have belonged, to this class. Their eruptive periods are likely to begin with explosive activity, manifested by the ejection of gases in great quantity, accompanied by solid fragmental material — bombs and ashes. In a succeeding phase liquid material issues; either it is ejected by the still issuing gases or it breaks through the crater walls and produces outflows of lava, sometimes of great volume. Finally the volcano becomes quiet, its energy for the time being exhausted; the lava column sinks down in the conduit and a period of quiescence intervenes before the next eruption. Although this sketch gives in a general way the succession of events, it must not be supposed that all volcanoes of this class are alike in the character of their eruptions, or that the same one always

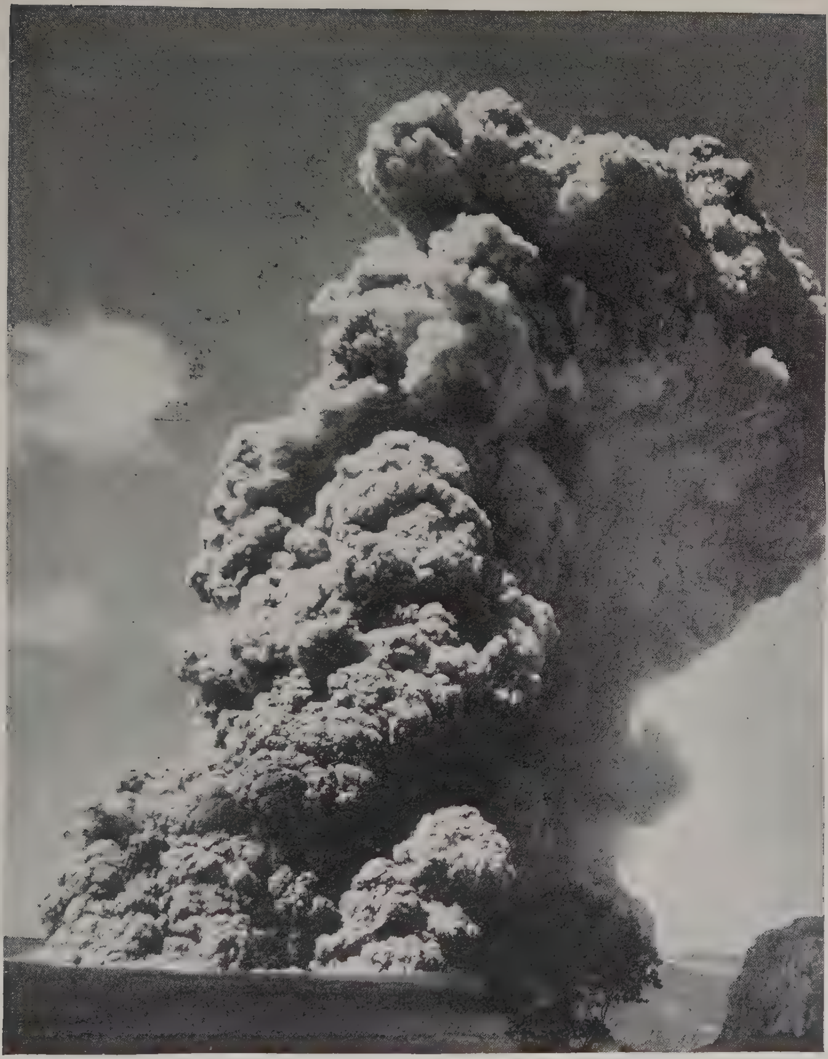


Fig. 155. — Eruption of Halemaumau, Kilauea Volcano, Hawaii, May 24, 1924. This "cauliflower cloud" was considerably more than a mile high when the photograph was taken. Violent outbursts of this kind are rare in Hawaii; ordinarily the eruptions of Kilauea and Mauna Loa are of the quiet type. (Maehara, Hilo, Hawaii.)

passes through a similar set of phases at each eruption, for there is great variability in these respects. The main point is that volcanoes of the intermediate class at times are explosively active and at other times quietly emit flows of liquid lava.

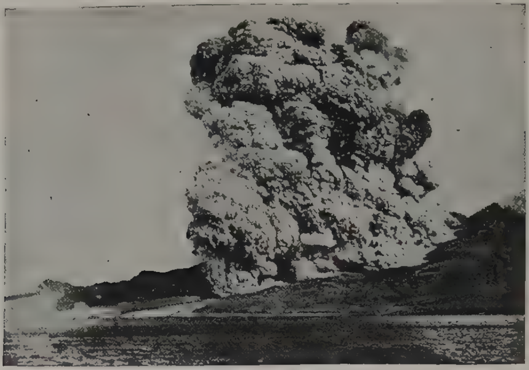


Fig. 156. — Fiery cloud of Mont Pelée descending the mountain slope to the sea. The cloud at this moment is 7000 feet high and moving forward at the rate of over a mile in 1.5 minutes. (A. Lacroix.)

Vesuvius, the longest and most studied and therefore the best known volcano in the world, belongs in this class. It occupies the site of an older volcano, which in the time of the Romans appeared to be extinct, for, although they recognized its nature, they had no traditions of its having been active. In the year A.D. 79 the volcano became active in eruptions that destroyed the towns of Herculaneum and Pompeii on its seaward flanks. A great part of the older cone on the side toward the sea was blown away or engulfed, and in its place the new center of activity, the modern Vesuvius, began to be built. This building up has continued until the present cone has become 4000 feet high. Partly enclosing it lies the sickle-shaped ridge of Monte Somma, the remains of the rim of the older crater (Fig. 157). Vesuvius is in a state of almost constant, relatively mild activity, with irregular periods of violent eruption. The last great eruption occurred in 1906.

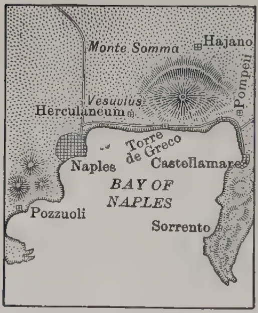


Fig. 157. — Map of Vesuvius and vicinity.

From the nature of the material composing their cones it is probable that the great volcanoes of the northwestern United States, previously mentioned, and now quiescent or extinct, belonged in the intermediate class, as well as the active volcanoes of the Alaska Peninsula and their extension on the Aleutian Archipelago.

Quiet Type. — Volcanoes of this type give rise to outflows of liquid lava, unaccompanied by the explosive disengagement of gases and the ejection of solid material as dust, ashes, and bombs. The lava of these eruptions is very hot and highly fluid. There is a more or less constant escape of gases from it, but without the catastrophic violence of the previous types. The best examples are in Hawaii.

The island of Hawaii consists of a vast mass of lavas surmounted by five cones: Kohala; Mauna Kea, now extinct (13,800 feet high); Hualalai (8300), active in 1801; Mauna Loa (13,700), now active and some of whose lava flows have been 50 miles long; and Kilauea (4000 feet), which is a great crater pit, rudely oval in shape and 9 miles in circumference, on the eastern slope of Mauna Loa and 20 miles from its summit. In the floor of this pit is Halemaumau, a circular depression, which before 1924 was 300 feet in diameter, and was occupied by a lake of liquid lava, boiling and fountaining from the escape of gases (Fig. 158). The temperature of the lava ranged from 1000° to 1200° C., the higher temperatures prevailing at times of increased activity. In 1924 the bottom of the pit in Halemaumau dropped suddenly 700 feet, and this subsidence was followed by explosive eruptions, the first since 1790. Avalanching from the sides of Halemaumau began and

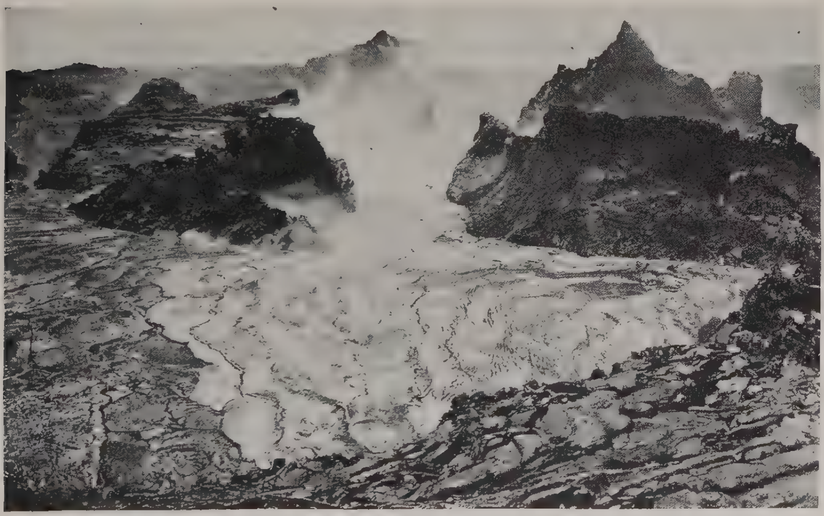


Fig. 158. — Halemaumau, north pool. October, 1921. (Hawaiian Volcano Observatory.)

enlarged the pit, so that it became 3000 by 3400 feet across and 1340 feet deep. After the eruption the pit filled again with lava. (Fig. 155.)

Mauna Loa is the " monarch among modern volcanoes." It exceeds all others in its mighty size and in the magnitude of its eruptive activity. At times immense columns of molten lava play as fountains several hundred feet high and afford a spectacle truly sublime. Its crater is at an elevation nearly 10,000 feet higher than that of Kilauea. The outflows of lava are more likely to occur through its flanks than through the crater rim, and sometimes they take place below sea level.

Relation between Volcanoes and Magmas. — The molten rock-matter that originates within the Earth's interior and gives rise to volcanic action and volcanoes is known as *magma*. When this issues at the Earth's surface, the liquid material, and the rock produced by its cooling and solidification, is called *lava*. It must not be supposed, however, that the composition which a solidified lava shows, if determined chemically, is exactly the same as that of the magma that yielded the lava. For the deep-seated magmas contain, in addition to the mineral substances of lavas, great quantities of gases, especially water vapor, which are dissolved in them under pressure. As the magma rises to the Earth's surface and the pressure on it is consequently diminished, the gases escape, usually with more or less explosive energy, and give rise to the spectacular features of volcanic activity. As the different types of volcanoes and of the lavas that they yield depend in large measure on the magmas producing them, it is necessary at this point to consider the nature and composition of these magmas.

Composition of Magmas. — As indicated above, the substances that compose a magma may be divided into two classes: 1, those which when hot are *volatile* and which mostly escape as gases and vapors, such as steam, carbon dioxide, hydrochloric acid, sulphurous vapors, during the congealing of the lava; 2, those which are *non-volatile* and remain to form the essential ingredients of the solid lavas. These fixed constituents are *silica* (SiO_2) and the oxides of six metals, *aluminum, iron, magnesium, calcium, sodium,* and *potassium.*

Silica is an important constituent of all magmas, and metallic oxides occur in all of them, but the particular metallic oxides present in different magmas range from almost nothing to considerable quantities. However, a kind of general rule governs the composition of magmas; without going into details, which are given in the chapter on igneous rocks, it may be said that the magmas, although forming a continuous chemical series, can be divided into two classes: one in which silica and the alkali metal oxides — soda and potassa (Na_2O and K_2O) — predominate, and the other in which, conversely, lime (CaO), iron oxides

(FeO and Fe_2O_3), and magnesia (MgO) predominate. Magmas of the first class, on cooling slowly, crystallize into a mass of mineral grains that consist chiefly of alkalic feldspar,[1] with which quartz is commonly associated; the resultant rocks are called silicous lavas. Rhyolite is the most abundant of the siliceous lavas. Magmas of the second class yield on cooling little or no alkalic feldspar, but abundant lime, iron, and magnesia minerals, such as pyroxene, calcium feldspar, and magnetite; rocks of this kind are termed basic lavas. As a rule the siliceous lavas are light-colored, whereas the basic lavas are very dark or black, and heavy because of their content of iron minerals. The siliceous lavas are sometimes termed *acidic* lavas because of the predominance of the acid-forming radicle (SiO_2) in them; the *basic* lavas are so termed because of the predominance of the bases (lime, iron, and magnesia) in them. Basalt is by far the most common of the basic lavas.

Relation to Volcanic Eruption. — The siliceous lavas, or rather the magmas that produce them, are, when their volatile constituents have escaped, thick viscous liquids, even at very high temperatures, as high as 2000° C. as experimentally ascertained. Parenthetically, it may be remarked that 2000° is far above the temperatures that prevail at existing volcanic vents, which are generally between 1000° and 1100° and rarely reach 1200°. The extreme viscosity is due chiefly to the high percentage of silica they contain, the amount in some kinds being as much as 75 per cent of the whole. For this reason when siliceous magma rises into the upper part of the conduit where the pressure is small and the contained gases begin to escape, the magma becomes stiffly viscous and the remaining gases can escape only with difficulty and usually with violence, giving rise to explosive eruptions. Hence volcanoes that yield siliceous lavas are likely to be of the explosive type, as Mont Pelée. On the other hand the basaltic magmas, with about 50 per cent of silica, are very much more fluid, and they remain quite fluid down to much lower temperatures, probably down to 1000° C. or lower: the gases escape from them readily, but without explosive violence, as illustrated in the lava lake of Kilauea in Hawaii. Hence the basaltic magmas usually cause quietly eruptive volcanoes.

The above statement indicates the general rule; it does not mean that basaltic volcanoes never have explosive eruptions, for a basaltic magma may become cooled in the conduit and in consequence be viscous, and thus permit its magmatic gases to escape only with difficulty and explosive energy (Fig. 155). The explanation applies chiefly to the two extremes and indicates what is probably the most effective cause for the explosive and quiet types of volcanoes. The intermediate type of volcano may be due in part to the intermediate kind of magma erupted from it, or to this factor combined with variations in viscosity at different stages during the eruption as well as variations in the amount of gases. Sudden accession of ground water into the volcanic pipe, as undoubtedly occurred at Kilauea in 1924, will produce an explosive eruption even in a basaltic volcano.

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[1] For description of these and other minerals see Appendix A.

PRODUCTS OF VOLCANOES

Gases. — It has been shown already that the products yielded by volcanoes may be divided into three general classes: gases and vapors, solid fragmental material, and liquid lava. These products will be considered in more detail, beginning with the gaseous substances. The quantity of steam discharged by active volcanoes is immense, and is indicated by the height and volume of the cloud with which many eruptions begin. This cloud consists of the dust and ashes borne aloft by the uprushing column of gases. The great quantity of steam thus discharged into the atmosphere may give rise by condensation to heavy downpours of rain in the vicinity of the volcano; and, owing perhaps to the friction of the particles and to atmospheric disturbance, the eruptions and rains are accompanied by conspicuous electrical displays. As an instance of the quantity of water that may be discharged, Fouqué estimated that one of the subsidiary cones of Mount Etna discharged in 100 days in the form of steam the equivalent of over 460,000,000 gallons of water.

In addition to the water, the different gases and volatile products exhaled by volcanoes make a long list. Not only are they given off from the vent itself, but the outflows of lavas continue for weeks and even months after their extrusion to emit gases as they cool and harden. Carbon dioxide, hydrochloric acid, hydrofluoric acid, and even hydrogen are given off, and to the mixture of the latter with oxygen and its sudden combustion are sometimes ascribed the explosions in the conduit. Sublimed sulphur and various compounds of sulphur, such as sulphuretted hydrogen (H_2S) and sulphur dioxide (SO_2), are emitted by some, but not all, volcanoes.

Fragmental Products. — These are the volcanic projectiles, the materials blown into the air by the sudden liberation of the gases. They may be derived from the crust, or plug, of hardened lava left in the upper part of the conduit after a previous eruption, from rock material torn from its walls, or from lava ejected from the upper part of the liquid column by the violent escape of the gases from the magma. Although the lava may start on its aerial flight in a liquid condition, it generally hardens in its passage and falls in solid form. The pieces of rock and the particles of magma driven upward and solidified are of all dimensions: from dust so fine that it may float in the atmosphere for several years, to large masses of several tons in weight. According to size, they are roughly classified as follows: pieces the size of an apple, or larger, are called *blocks* if ejected as solid fragments and *bombs* if ejected as particles of still-fluid magma; those the size of a nut are termed *lapilli* (meaning little stones); those the size of a pea are volcanic *ashes*, while the finest is *volcanic dust*. The ashes and lapilli are frequently spoken of as *volcanic cinders*, and the cones made of them as cinder cones. It should be clearly remembered, however, that although these terms are used to describe the appearance of the products, the "cinders" are not products of combustion. A volcanic bomb is illustrated in Fig. 159; it is of the variety called bread-crust bomb.

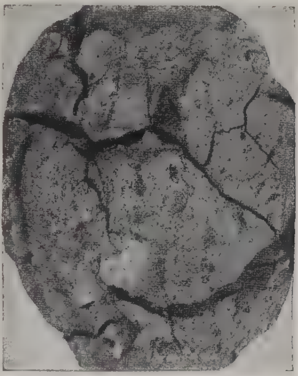
Fig. 159. — Volcanic bomb, Lipari Islands.

The ejected products are in part composed of compact solid rock and in part are of a spongy, cellular, or vesicular character. This vesicular condition is due to the fact that, while the major part of the gases is passing into the air and carrying the fragments with it, a minor part is expanding in the particles of liquid, puffing them up into the cellular forms. Although the bombs, lapilli, and most of the ashes fall in the immediate vicinity of the vent and thus help to build up the cone, the dust may be carried by the prevailing winds long distances, hundreds of miles or more, and be thus spread over an immense area. Huge quantities are discharged in great eruptions, amounting to many millions of tons. Such dust showers generally are very destructive to vegetation and even to animal life, but the soil ultimately yielded by them is very fertile.

Liquid Materials; Lavas. — In volcanoes whose periods of eruption begin explosively, the liquid lava generally issues later, after the vent has been cleared. The volcanic cone is not a structure of great strength, and is easily ruptured, or fissured, by the explosions and the pressure of the lava column, and hence the magma is not likely to flow out over the lip of the crater, but to issue through fissures in the sides of the cone. It may even happen, especially when the cone is composed of cinders, that, unable to withstand the pressure, one side gives way and allows a flood of lava to rush out through the breach thus made.

The appearance and character of a lava stream and the material produced by its solidifying depend on several things: on the chemical nature of the magma, on its viscosity, and the extent to which it retains its dissolved gases. On the chemical composition will depend the nature of the resultant rock, whether it will be a light-colored lava or a black basalt, or of intermediate character as previously explained. On the viscosity will depend the rate at which the lava will flow, the distance to which it will flow and in large measure the appearance its surface will present. When it issues, the lava is red or even white hot. It soon cools on the surface, darkens, and crusts over. If very viscous, the under part may yet be in motion, and the crust breaks up into a mass of rough, angular, jagged blocks of rock, which are borne as a tumbling, jostling mass on the surface of the slowly-moving flow. When eventually the flow comes to rest, the lava sheet is extremely rough and difficult to traverse. Such lava flows in Hawaii are called *aa* by the natives.

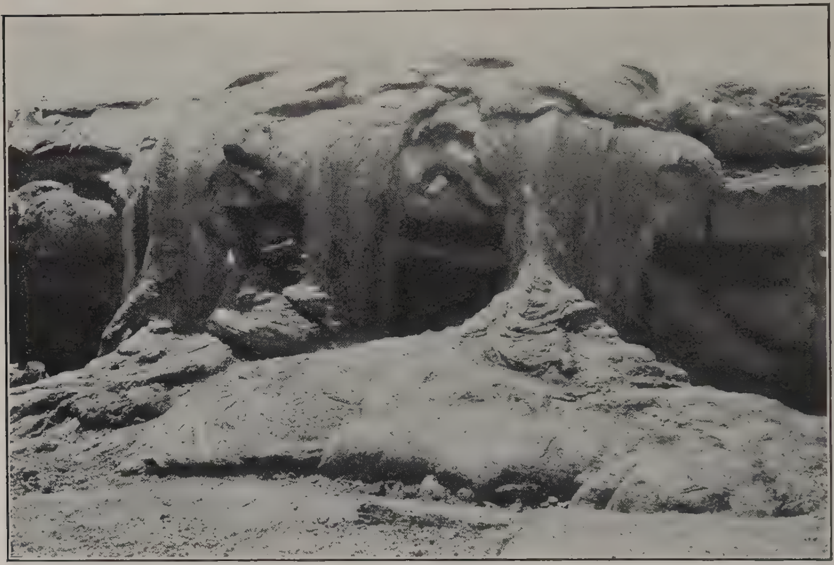

Fig. 160. — Lava cascades, Hawaii. Although the lava is now cold and solid, its form still suggests the fluid condition in which it poured over the cliff. This illustrates the smooth pahoehoe type of lava. (U. S. Geol. Surv.)

In marked contrast other lavas may harden with smooth surfaces, which exhibit curious ropy, curved, wrinkled, or twisted and billowy forms, as seen in Fig. 160. Lava of this kind the Hawaiians term *pahoehoe*, in reference to its glistening, satiny surfaces. The difference between the two varieties of lava is determined in some way yet unknown by differences in the physical conditions during consolidation: for a flow may begin as pahoehoe and end as aa.

Sometimes on slopes, after the lava has crusted over, the still liquid portion beneath runs out at the lower end, leaving long galleries, tunnels, or caves beneath the crust. On some volcanic cones the natural drainage passes into these tunnels, disappears from view, and issues lower on the slopes in the form of springs. Such may be in part the cause of the springs around Mt. Shasta.

Some lavas when ejected are too viscous to flow; they may then pile up over the vents in great domes. Such doming is chiefly, if not wholly, confined to the siliceous varieties of lava. Domes of lava in central France, Bohemia, and Germany are thought to have been formed in this way. After the violent eruptions of Pelée, in 1902, the column of siliceous lava that filled the vent and had hardened into rock was pushed up so that it rose like a vast tower above the volcano, until it attained a maximum height of 1000 feet (Fig. 161). Gradually it crumbled into a mass of blocks as a result of continuous explosions of gases.

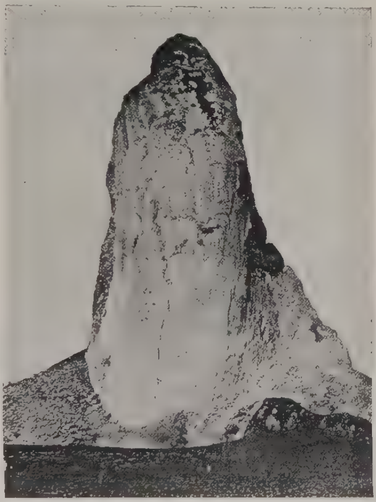

Fig. 161. — Rock tower of Mont Pelée, Martinique. 1000 feet high. (A. Lacroix.)

Effect of Contained Gases; Vesicular Lava. — That the lavas, even after they issue from the volcanic vent, still contain dissolved gases is abundantly shown not only by the clouds of steam that issue from them but also by the structures they assume as they cool into stone. Thus the upper part of a flow, especially of viscous lavas of the siliceous class, may be so puffed up by the innumerable bubbles of vapor in it expanding on relief of pressure that it may become a veritable glassy froth. Such rock froth, which is usually white or light-colored, is known as *pumice*, or *pumice stone*.

In more fluid lavas, especially those of the basalt class, the bubbles are larger, and the resultant rock is spongy, cellular, or vesicular. This porous, cindery, or slag-like form of lava is called volcanic *scoria*. It is usually dark to black, or reddish (Fig. 162). Pumice, scoria, and other vesicular products are characteristic features of the upper surface of lava flows, and they constitute also a major part of the coarser fragmental materials, such as bombs and lapilli, that build up the volcanic cone.

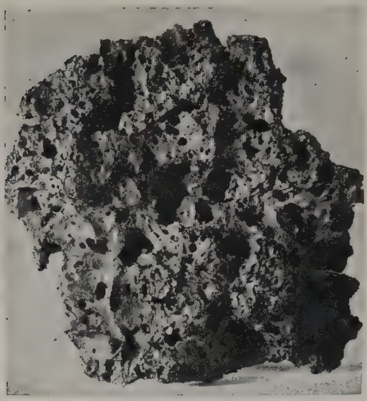

Fig. 162. — Volcanic scoria.

Consolidation of Lavas; Glass and Stone. — After lavas have been poured out and have solidified, most of them present the ordinary appearance of *stone*, but some, instead, have that of *glass*. The reason for this is as follows. If the liquid lava is not too viscous, the chemical molecules composing it will be capable of motion and will arrange themselves into definite compounds; that is, will crystallize into mineral grains, or crystals. Possibly the crystal grains thus formed will be large enough to be seen readily and the constituent minerals will then be determinable, or they may be so minute that the lava has a homogeneous appearance; nevertheless if crystallization has taken place the lava has the aspect of stone. On the other hand, if the lava is extremely viscous or quickly becomes so through rapid cooling, the molecules may not be able to arrange themselves, or crystallize, into minerals, and the mass solidifies as a glass.

Glasses form chiefly as the result of the freezing of siliceous lavas for, as previously explained, they are usually the more viscous. Volcanic glass is called *obsidian;* certain varieties, in allusion to their luster and appearance, are called *pitchstone.* Some obsidians are pure glasses, others are mixtures of glass and crystals. In Yellowstone Park, Obsidian Cliff presents a section of volcanic glass 100 feet thick, which has cracked into columns in cooling. Such a thickness of purely glassy lava is unusual. It is chiefly on the edges and upper surface of lava streams that these glassy forms are found. Primitive peoples, before they gained a knowledge of metals, made much use of obsidian for knives and for arrow and spear points.

VARIETIES OF VOLCANIC CONES AND CRATERS

Kinds of Cones. — The nature of a volcanic cone depends on the material of which it is built. If composed wholly of fragmental products, the cone is high and steep in proportion to its size. This steepness is due to the high angle of repose for lapilli and volcanic ash; such material is very angular, rough, and clinging, and slopes of 40° are attained before the accumulating mass begins to slide. Cones of this kind are called *cinder cones,* and they are characteristic of volcanoes of the explosive class (Fig. 163). In contrast with them are the *lava cones* (Fig. 164), formed entirely by quietly outflowing liquid lavas, like that of Mauna Loa. They are necessarily very low and flat in proportion to their size, the angle of inclination being less than 10°. Most volcanoes, however, and this includes most of the largest volcanoes in the world, are of the intermediate type in their eruptive activity, and in consequence their cones have forms that are intermediate between those just described. For they are built up by the fall of ashes and lapilli when they are explosively active and by lava flows when they erupt quietly. The great cones of the Pacific States — Mts. Shasta, Hood, Rainier, and others of the Cascade Range — have been built in this way.

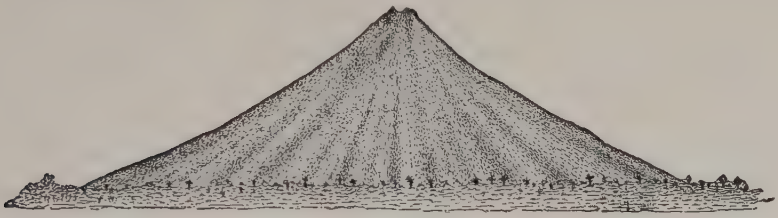

Fig. 163. — Cinder cone, showing steep angle of repose of lapilli. Outline as given by a photograph of Mayon volcano in the Philippines.

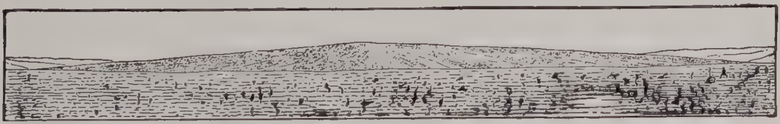

Fig. 164. — A lava cone, to show contrast with Fig. 163. From the Snake River plain, Idaho.

The eruptions that break out on the lower flanks of the larger volcanoes give rise to smaller, or "parasitic" cones. Mt. Etna is surrounded by over 200 of these, some of which are nearly 700 feet high. San Francisco Mountain in Arizona, an extinct and partly eroded volcano, has a number of such minor cones, some of them remarkably well preserved. As an active volcano grows, the earlier parasitic cones may be buried and concealed under later accumulations, or, in the declining stages of activity, the eruptive energy may show its last efforts in building them, as appears to have happened at the San Francisco volcano, just mentioned.

Calderas; Explosion and Subsidence Basins. — The term caldera, from the Spanish for caldron, is applied to crater-like basins of great size, especially those which are very broad as compared to their depth. The name is taken from the huge pit in the Canary Islands, called La Caldera, which is from 3 to 4 miles wide and bounded by lofty cliffs 1500 to 2500 feet high, except on one side where the encircling wall is breached. From without, as seen from a distance, the general aspec- of La Caldera is that of a huge cone truncated far below its apex.

Many such great calderas occur in various parts of the world. Thus Tamboro in its explosive eruption of 1815, previously alluded to, blew away a good part of the original volcano and produced a caldera nearly 4 miles in diameter. Perhaps more generally the caldera has been produced by the subsidence of the column of liquid lava in the conduit of the volcano, leaving a great cavity into which the superstructure of the cone has subsided.

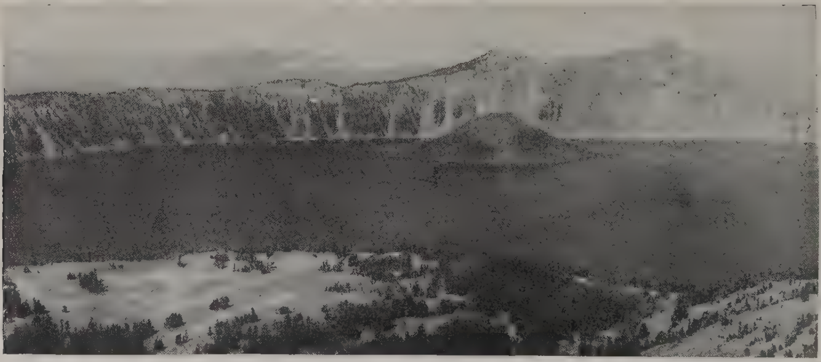

Fig. 165. — Part of the basin and wall of Crater Lake, Oregon. Note the small cone within (Wizard Island). (U. S. Geol. Surv.)

The finest caldera in the United States is at Crater Lake in southern Oregon. This marvelously beautiful lake occupies a caldera at the summit of a volcanic mountain in the Cascade Range, and is 6 miles long by 4 broad, 2000 feet deep, and encircled by steep cliffs 500 to 2000 feet high (Fig. 165). An island in it, made by a small but perfect cone of volcanic material, indicates a feeble renewal of eruptive activity after the principal subsidence. The caldera, if emptied of its water, would appear as a great basin. The reason for believing that the caldera was formed by the collapse and engulfment of the greater part of a former cone as the result of the subsidence of the lava column in the pipe of the volcano, rather than by explosion, lies in the absence of the débris — about 18 cubic miles of material — that so gigantic an explosion would have spread over the adjacent outer slopes. The former mountain, to which the name Mt. Mazama has been given, is conceived to have had the size and general character of Mt. Shasta and to have been heavily capped with snow and glaciers during the Glacial Period.

Explosion Pits. — In some places where volcanic activity has begun it has proceeded no further than the initial explosions that drilled a vent through the country rock. The material blown out has made a low ridge around the pit, but no real cone was built. Sometimes volcanic products, such as pumice and cinders, are mixed with the fragments of the country rocks. Such basins range from a few hundred feet to several miles in width, and in humid regions they are usually filled with water and form lakes. Some of the best examples of them are in the region west of the Rhine in Germany, known as the volcanic Eifel. They are called *maars* (German, *maaren*), like the Pulvermaar. A pit that strongly resembles a maar exists at Coon Butte in Arizona. The basin sunk in the plain is $\frac{3}{4}$ of a mile in diameter and 500 feet deep. The presence of meteoritic iron in and about it and other features have led to the view that it was caused by the impact of a huge meteorite and is probably not of volcanic origin. Hence it has recently been renamed Meteor Butte.

Structure and Dissection of Volcanic Cones

Structure of a Composite Cone. — If a column of magma is forced upward through the crust of the Earth until it reaches the surface, the relief of pressure will enable it to commence discharging its dissolved gases and vapors. Conceivably the pressure of the contained gases may be too great for the topmost layers of the bedrock to restrain them until the magma reaches the surface; consequently these layers may be blown into the air and a vent drilled ahead of the rising column of lava. Arrived at the surface, the magma may flow out quietly, or, if it is too viscous to do this, explosions may continue and material be blown upward. The pieces cannot, of course, fall back against the uprushing column of gases and cover the vent; they must fall outside of it, the heaviest

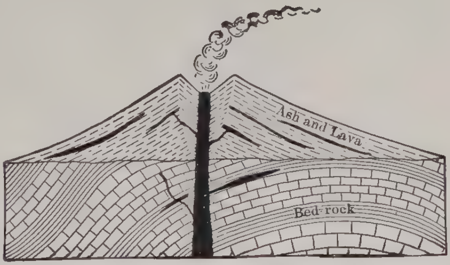

Fig. 166. — Ideal section through a volcano. The dark layers in the cones are either buried flows or injected masses (dikes and sills).

and largest first and nearest to it, the smaller and lighter later and farther away, the distribution of the lighter depending much on the wind. Thus the cone grows as a circular ridge upon whose crest most of the ejected material is deposited. This material tends to roll and slide both outwardly away from the center and inwardly down the crater toward the vent. This process forms the cone and crater, and certain features of their structure follow as a consequence of this mode of formation (Fig. 166).

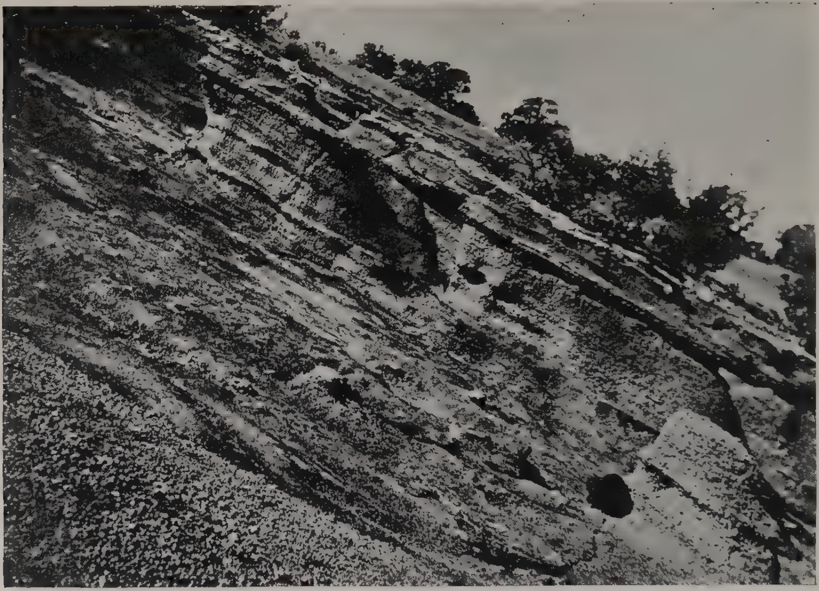

Fig. 167. — Inclined beds of volcanic ash. Part of a former cone at Trinchera, Colo. (U. S. Geol. Surv.)

Tuff and Breccia. — The deposits of successive eruptions are marked by layers, some of coarser, some of finer material, in each of which, if not composed of uniform-sized particles, there is a gradation from coarser at the bottom to finer at the top. Thus there arises a rude stratification, or bedding, the beds sloping down and out from the crater edge (Fig. 167). The bombs, lapilli, and ash composing the beds gradually become compacted by their weight and by the infiltration and deposition of cementing substances. They are thus transformed into a more or less friable, porous rock called, when composed of the coarser materials, *volcanic breccia,* and when of the finer dust and ashes, *volcanic tuff.* In the crater the fragments are larger, generally large blocks of rock, and they usually form a tumultuous mass without order or arrangement, with intermingled finer material; such material is called *volcanic agglomerate.*

Lava Flows and Dikes. — In addition to the beds of tuff and breccia, liquid lava flows down the outer slopes of many volcanic cones, and when these streams harden into solid rock they protect the softer layers of tuff and breccia from erosion and give strength to the edifices. Since the lava rarely flows over the lip of the crater, but, especially in high cones, breaks through fissures on the sides of the cone, these fissures also become filled with lava, which hardens into rock. These rock-filled fissures are called *dikes,* and like ribs they also serve to strengthen the volcanic cone. Thus a vertical section through a volcano (Fig. 166) shows a central core of igneous rock surrounded by beds of tuff and breccia with intercalated flows of lava, which are cut by a radial system of dikes. This description gives an idea of the general structure of a typical composite cone, one formed by the intermediate type of volcano.

Dissection of Volcanoes. — At every stage of its existence a volcano is subject to the agencies of erosion and weathering, which tend to cut down all prominences on the Earth's surface. Its height and appearance at any given time are the result of the balance between these destructive forces and the upbuilding power of volcanism. Even active and growing volcanoes commonly are trenched and scored by ravines and gulches. After eruptions, when the cones are covered with fresh deposits of dust and ashes, the latter become so saturated with water from the rainfall that they slide down as flows of liquid mud, leaving ravines, which are enlarged by subsequent storms (Fig. 168).

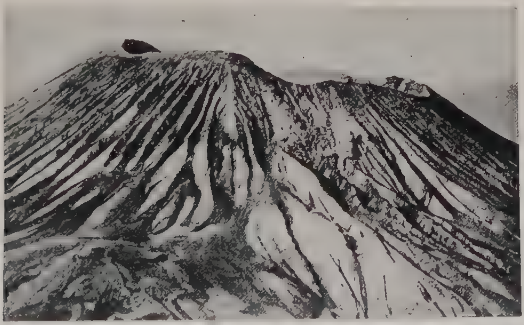

Fig. 168. — Vesuvius in 1906, showing trenching by ravines in the ashes after the great eruption. (F. A. Perret. Courtesy of Harper's Weekly.)

As soon as a volcano becomes extinct, the ravages of erosion are unchecked, and the period of dissection ensues. The lighter tuffs and breccias are carried away more easily and rapidly; the harder, more compact, and resistant flows and dikes of lava, and the parts protected by them, are eroded more difficultly and slowly. It is surprising, however, for how long a time some cones that are composed of mere cinders loosely piled will resist erosion and retain their form. The reason for this resistance to erosion is due to the porosity of the material, which allows the rainfall to sink through it without causing downwash.

As erosion progresses, the mass of rock formed by the solidification of the magma in the central conduit is brought to view, provided the magma column was not drained off before the volcano became extinct. If the central column of magma was drained off, the site of the vent probably is marked by a mass of agglomerate. As erosion progresses and the cone is demolished, the central rock mass, owing to its greater resistance, is likely to form a decided prominence, and, even when erosion has finally swept away all external evidence of the cone and bitten deeply into the underlying rocks, it may remain projecting, a monument to the vanished volcano. A rock mass of this kind filling a former conduit is termed a *volcanic neck* (Figs. 169 and 181).

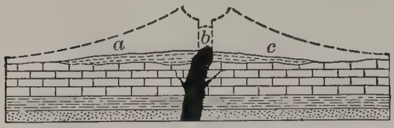

Fig. 169. — Section through a partly eroded volcano, with volcanic neck (*b*) left projecting. Broken lines show ideal restoration of cone and conduit.

Volcanoes and Deep Masses of Magma. — In the preceding paragraph there has been sketched the structure of the volcano down to a conduit filled with magma derived from below. We cannot leave the subject of volcanic activity and the structures it gives rise to before stating that volcanoes are only one phase, the surface manifestation, of the general movement of masses of magma from unknown depths into the upper region of the Earth's crust. Although these masses of magma may attain the surface and there produce volcanoes or lava flows, many of them have not reached the surface, but have remained in depth intruded into the rock layers of the crust, and there cooling and solidifying, gave rise to rock bodies of varied shapes and of sizes from a few feet in thickness up to miles in extent. Such deep bodies of magma form what are known as intrusive masses of igneous rock, and every extinct volcanic conduit, if it could be traced downward, would be found to join some such intrusive mass below, or, if active, to extend into a body of magma which in time will solidify as an intrusive mass. A proper understanding of the intrusive masses demands a more extended description and explanation, which will be found in the next chapter.

Life and Distribution of Volcanoes

Age of Volcanoes. — The span of life of active volcanoes differs greatly in different individuals. We know from written testimony that Etna has had the same general eruptive character for the last 2500 years. We estimate that because of its great volume it has taken at least 300,000 years to build this grand volcano. It is certain that the eruption of such vast masses of material as make up the larger volcanoes must have required, from the human standpoint, an immense lapse of time. On the other hand, we know from various considerations that the present active volcanoes are from the geological standpoint recent affairs.

It is also difficult to say whether a volcano is extinct, because long periods, hundreds of years, may elapse between eruptions. In the Middle Ages, Vesuvius had been so long dormant that its crater was overgrown with vegetation and it gave no sign of life. But in 1631 it became violently eruptive, and has since been intermittently active.

New Volcanoes. — Within the period of recorded human knowledge a number of volcanoes have begun their existence, and many of them are still active. Jorullo, in Mexico, came into being Sept. 28, 1759, in the midst of a cultivated plain, and is now about 4300 feet high; Izalco in Salvador, which began in 1770, has been almost continuously active since then, and is now over 6000 feet high.

No well authenticated volcanic eruption was witnessed within the limits of the United States proper until May, 1914, when explosive eruptions, which have since continued at intervals, began at Lassen Peak in northern California. The eruptions have been chiefly of gases, ashes, and stones. In 1915 there were two enormous explosions — two horizontal blasts that laid waste a wide swath of country.

Distribution of Volcanoes. — The present active vents are about 430 in number, and those cones which because of their slightly eroded condition may be considered dormant or only recently extinct amount to several thousand. They have a general tendency to be grouped in long belts on the Earth's surface. The most marked of these belts is the great zone that borders the Pacific Ocean; it passes northward along the Andes, through Central America into Mexico, through the United States and Canada to Alaska, then along the Aleutian chain to Asia, and turning southward through Kamchatka, Japan, and the Philippines, it crosses the East Indies, and by various island chains again passes into the Pacific. Certain portions of this belt, like the Andes and the Aleutian chain, are remarkably linear and well developed. Another great belt has an east and west direction: from Central America it extends through the West Indies; it then continues through the

Atlantic by the Azores, Cape Verde, and Canary islands, runs through the Mediterranean, through Asia Minor and Arabia, and continues along the chain of the East Indies, where it crosses the circum-Pacific belt, and extends out into the Pacific. There are some isolated groups with irregular distribution. (Figs. 170 and 171.)

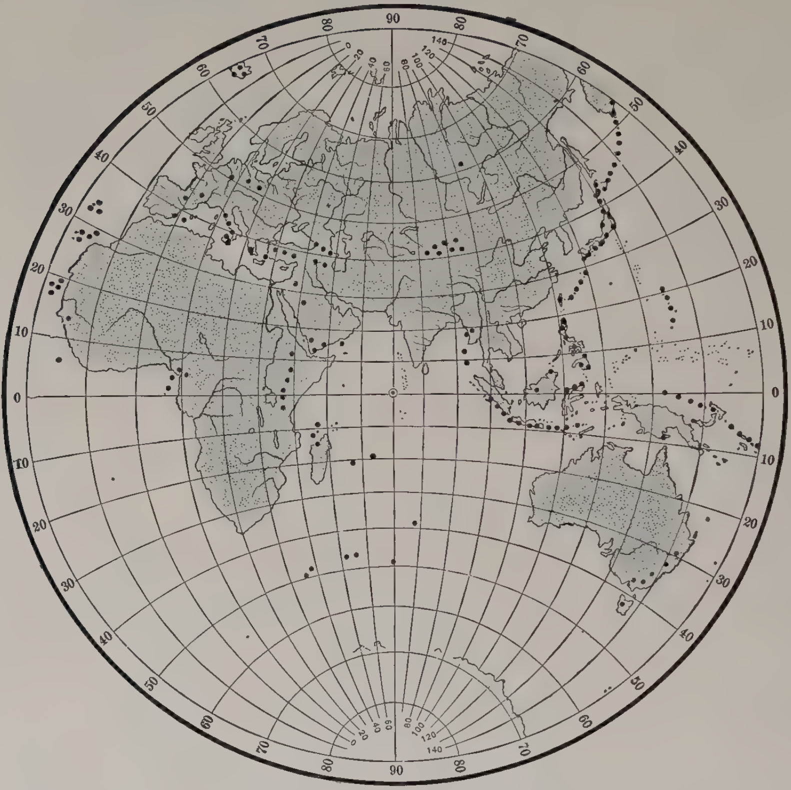

Fig. 170. — Map showing distribution of active or recently extinct volcanoes in the Eastern Hemisphere. On S. L. Penfield's stereographic projection.

Volcanoes occur both on the continents and in the oceans, the true oceanic islands appearing to be entirely volcanic. Notably in the Pacific there are great numbers of them, many extinct or dormant, some still active, and here again many of the volcanoes are grouped in lines and stand on the submarine ridges which rise from the ocean floor. From the fact of linear arrangement has been drawn the important deduction that volcanoes are in general situated on, or near, lines of fracture, folding, and weakness in the Earth's crust.

Lines of fracture and weakness have undoubtedly proved favorable sites for volcanic action, not only for a time, but in places for long-continuing geologic periods, and thus they have greatly influenced the origin, situation, and arrangement of volcanoes. But, on the other hand, it seems clear that a volcano, or a group of them, may originate where no definite connection between them and any fracture line can be shown to exist. And in places no tendency to a linear arrangement in the group may be seen. The volcanic forces appear to have been sufficiently powerful to find an outlet without needing the aid of a fracture.

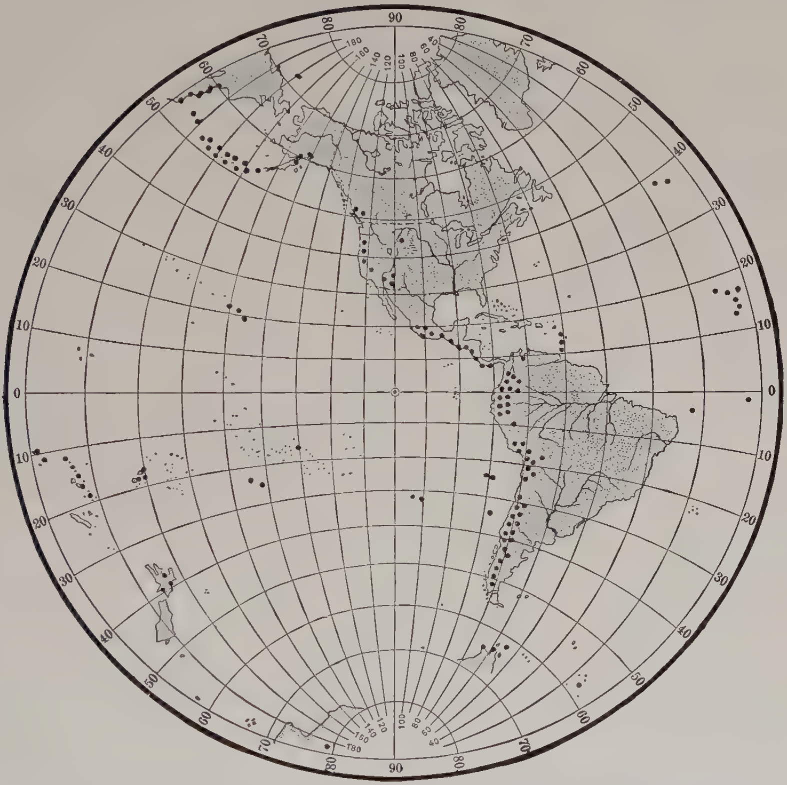

Fig. 171. — Map showing distribution of active or recently extinct volcanoes in the Western Hemisphere. On S. L. Penfield's stereographic projection.

A striking instance of how little influence surface topography may have in determining the site of volcanic action, which in the immensity of its power appears to disregard such minor considerations entirely, can be seen at the Grand Canyon of the Colorado. Uninfluenced by its 5000 feet or more of depth, volcanoes have broken out upon its very rim, instead of in its depths, and their lavas have flowed down into it.

Submarine Eruptions. — From the great number of volcanic islands in the sea, the real oceanic islands being of this nature, it is evident that in times past tremendous eruptions and vast outpourings of lava have occurred on the sea floor. The volcanic chain of the Hawaiian Islands is an example of this. Actual eruptions beneath the sea have been observed and recognized by the issuance of vapors and ashes from the water. Thus, in 1831 a volcano was thrown up in the midst of the Mediterranean Sea, forming a new island called Graham's Island. Being composed of light cinders, it was soon destroyed by the waves and reduced to a shoal. The three Bogoslov volcanoes of the Aleutian chain formed in 1796, 1883, and 1906 are other examples.

Fissure Eruptions. — Outflows of basaltic lava have taken place in several regions on such a gigantic scale and deluged such immense tracts of country that they cannot be referred to the outpourings of any volcanic cone or group of volcanoes. Moreover, the cones from which they might have come are apparently wanting. These great lava floods have issued from fissures in the Earth's crust. The result of such flooding is that broad plains, or plateaus, consisting of successive level sheets of basalt lava, in places interlaid with beds of tuff, have been formed.

Basalt plateaus of this origin are the great lava fields of the Columbia and Snake rivers in the far Northwest of the United States, which cover from 200,000 to 250,000 square miles and are in places 4000 feet thick; the Deccan traps (basalts) of western India, which are at least 200,000 square miles in extent and reach a maximum thickness of 6000 feet; the northern British Isles, which in part, with the outlying island groups, appear to be remnants of a great basalt plateau that may have extended to Iceland. The horizontal layering of these lava sheets is evidently due to the extreme fluidity of the issuing magma, which permitted it to flow for many miles before congealing. Thus in Iceland, a lava flow has been traced a distance of 60 miles. It is such outpourings, which occur in other regions as well as in those mentioned, that exhibit to us the grandest effects of volcanism. It is conservative to say that since a comparatively recent geologic period as much as 500,000 cubic miles of molten material have been transferred from the inside to the outside of the globe by the extrusive process of volcanism, most of it by fissure eruptions.

Origin of Volcanoes

General Remarks. — In considering the nature of volcanoes, the character of their eruptions and of the products afforded by them, their distribution, and in some measure their life, we are dealing with ascertained facts. We know also quite clearly the reason for the different kinds of eruption and the three types of cones. But when we seek to learn the cause and origin of volcanism we must then consider the depths of the Earth itself, about which we know very little. We are led from facts into almost pure speculation, and this should be clearly understood by the student. There are, however, certain phases of the problem of volcanism that may be considered here.

Problems of Volcanism. — Some important questions that arise when we seek to discover the cause of volcanism may be stated as follows: What is the origin of the *heat* that keeps the magmas in a molten condition? What is the origin and history of the *magmas* that come to the surface? From how deep down do these magmas come, and where is the seat of volcanism? What is the origin of the *gases;* have they always been contained in the magma, or have they been absorbed from outside sources, and, if so, when and where? And finally, what causes the magma to *ascend* to the surface from the depths and thus give rise to volcanoes? These are fundamental problems, most of which cannot be solved until our knowledge of the Earth's interior is increased considerably. Nevertheless, the views held regarding them may be briefly stated and discussed.

Origin of the Heat. — There is a persistent view that the heat necessary to produce the magmas is original, residual from a globe once molten and still intensely hot in its interior. Some, however, regard the heat as due to the gradual contraction and compression of the Earth through the force of gravity.

Since the discovery of radioactivity, a more promising theory has been developed. The rocks of the crust contain the elements uranium and thorium, which are spontaneously breaking down into other substances, with the production of heat in notable quantity. It is suggested that the changes of this nature that are going on within the Earth produce the heat necessary for volcanic action and even for such vast-scale igneous manifestations as are represented by the enormous intrusive bodies beneath the surface.

Origin of the Magma. — This is a complex problem. Different volcanoes erupt lavas of different kinds. Vesuvius erupts one kind, Etna another. Furthermore, many a long-lived volcano has erupted a variety of lavas. For example, Lassen Peak, which is regarded as an old and dying volcano, has erupted rhyolite, dacite, andesite, and basalt. Did the supply pipe of Lassen Peak tap four different reservoirs containing four different magmas? It is believed that originally there was a homogeneous magma beneath Lassen Peak and that by a series of internal changes this magma yielded the other magmas. This process by which a magma of initially homogeneous composition splits up into unlike fractions is called magmatic differentiation. It is to this process that appeal is made to account for the diversity of igneous rocks, not only at Lassen Peak but at all other volcanic centers.

Origin of the Gases. — The chief magmatic gas is water, as a rule making up more than 80 per cent of the total. This water may have been part of the original substance of the Earth, in short, of primitive origin — " telluric water " (from Tellus, the Earth) — or formed from the combustion of primitive hydrogen, or it may have been atmospheric water absorbed by the magma from the surrounding rocks, or it may have been acquired by the magma by melting up rocks containing water-bearing minerals. A quantitative evaluation of these possibilities is beyond the present powers of science. A magma that has dissolved much limestone would become highly charged with carbon dioxide, which is one of the important volcanic gases. Such absorption of limestone would generate enormous pressures, and such local development of pressure, it has been ingeniously suggested, may have determined the sites of certain volcanic vents.

The fact that so many volcanoes are situated in, or near, the sea or lakes has been considered a strong indication that the gaseous water contained in the magma has been obtained from descending surface waters. But this argument when examined loses its force. The nearness of some volcanic chains to the sea, like those of North and South America, is only relative to the size of the continental masses. Actually they are long distances inland: in South America from 100 to 250 miles and this includes some cones still active — like Cotopaxi — which are not near any inland water body; in North America from 30 to 100 miles or more, and, although these are mostly extinct, it can yet be shown that when active there were no bodies of water near most of them.

Cause of Ascension. — The fact that the great volcanic chains are situated on those belts along which movement and disturbance of the crust have taken place is significant. For these belts are apparently zones of weakness in the crust, and have thus afforded favorable places for the upward movement of the magma and its escape to the surface. In times past great segments of the Earth's crust have moved up or down with respect to one another. It is noticeable in many places that where one of these great blocks, measuring hundreds or thousands of square miles in area, has sunk, this subsidence has been attended by uprise of magma, outflows of lava, and commonly by volcanic action. Examples of such crustal subsidence and concomitant volcanic activity are the depressed tracts that form the great Rift Valley of East Africa and the valley of the Rhine. The mechanism of ascension of the magma in this way is likened to the action of a hydraulic press.

The old idea that the Earth has a hot, liquid interior and that the downward pressure of the cold and solid crust collapsing on the shrinking nucleus forces this liquid out and thus gives rise to volcanoes has been completely disproved by a number of considerations and is no longer held. The independent eruptions of adjacent volcanoes in the same group, and the fact that the lava column in Mauna Loa stands 10,000 feet higher than that in Kilauea, only 20 miles away, are disproofs of this view.

FUMAROLES AND HOT SPRINGS

Introductory. — In the foregoing description of volcanoes it has been shown what an active rôle gases, especially superheated steam, play in their eruptions. But long after a volcano has ceased to be active and has passed into a dormant or dying stage these gases continue to issue from its crater, or from its flanks, or even from places in the surrounding country. In the same way thick beds of extruded lavas continue, often for years, to exhale steam and other vapors. And, as we shall show later, it has often happened that bodies of magma have penetrated into the outer shell of the Earth without attaining the surface or forming volcanoes, and in solidifying they, like the lavas, have given off quantities of gases, which work their way through fissures upward to the Earth's surface. It is now proposed to describe the phenomena produced at the surface by such emanations. They may appear as vapors or in liquid condition; the gaseous emanations may be considered under the general heading of *fumaroles,* the liquid emanations under *hot springs.*

Fumaroles. — This word, which is derived from a Latin verb meaning to smoke, is applied to fissures or holes in the rocks from which steam and other gases escape with more or less force. The " smoke " of the fumarole is thus mainly steam. Although steam predominates, generally forming 99 per cent of the total, other gases, such as carbon dioxide, hydrochloric acid, hydrogen sulphide, hydrogen, methane, and others also occur. Fumaroles that give off sulphurous vapors are termed *solfataras,* from the Italian word for sulphur.

In addition to the substances already mentioned, fumaroles carry certain metallic constituents such as iron, copper, and lead. These metals have been rendered volatile by the presence of chlorides and fluorides in the magma and consequently are able to leave the magma and escape into the surrounding rocks. As they approach the Earth's surface they begin to react with the other fumarolic gases and are deposited in the form of metallic minerals in the fissures through which the gases are streaming. Hematite is probably the commonest mineral formed in this way. During one of the eruptions of Vesuvius a fissure 3 feet wide was thus filled with hematite in a few days. Galena, the chief ore mineral of lead, is occasionally formed at Vesuvius by the mutual action of the lead chloride and hydrogen sulphide contained in the fumaroles. Thus ores are actually being deposited under our eyes; in fact, it was the contemplation of these phenomena on the flanks of Vesuvius that first suggested the fruitful idea that there is an intimate relation between igneous rocks and the occurrence and origin of ore deposits throughout the world.

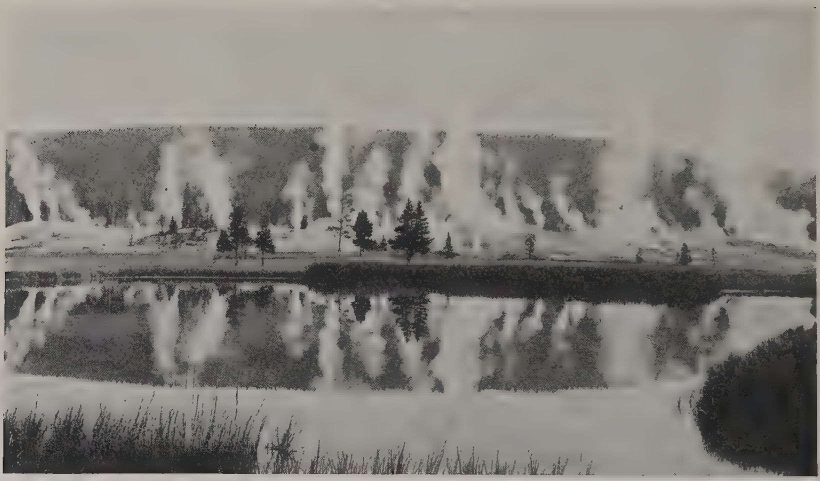

Fig. 172. — Beehive group of geysers, Upper Geyser Basin, Yellowstone National Park. (U. S. Geol. Surv.)

The volcanic cone near Naples known as Solfatara last erupted in 1198; since then it has been merely discharging steam mingled with sulphur vapor, and this has given rise to the use of the term *solfataric stage* when volcanoes become quiescent or are dying. Some of the great cones of the Cascade Range, like Mt. Shasta, appear to be in a solfataric stage. In Yellowstone Park the solfataric condition still prevails and fumaroles abound. Although the steam given off in fumaroles can be mostly ascribed to magmatic origin, the amount is often increased by descending surface water that becomes vaporized, either by contact with hot rocks or by volcanic exhalations, which, as already mentioned, are chiefly superheated steam. This is probably the case in Yellowstone Park (Fig. 172).

Carbon dioxide gas is given off copiously in many places where volcanic activity still abounds, and in many where volcanism has long since died out. In some places the carbon dioxide issues directly from the ground as a gas spring, and such occurrences are known as *mofettes.* Being heavier than air, in still weather it may collect in depressions near the vent, and, as it is colorless, tasteless, and odorless, such pools of gas are deadly traps for animals by suffocating the creatures that enter them. This is illustrated by " Death Gulch " in Yellowstone Park, where animals as large as grizzly bears have become asphyxiated. But the carbon dioxide is far more likely to encounter ground water on its way upward and thus give rise to a carbonated spring, which if it passes through limestone will dissolve some of the limestone and is therefore likely to deposit carbonate at the Earth's surface. Some carbonated springs, however, derive their carbon dioxide from the action of acid on limestone.

Hot Springs

Hot springs as well as fumaroles are likely to occur in volcanic regions. There is an intimate relation between hot springs and fumaroles: in many regions as the dry season comes on some of the hot springs become fumaroles, and when the wet season returns the fumaroles become hot springs. The evident seasonal variation leads to the theory that hot springs are chiefly fed by ground water that has become heated by magmatic steam. Generally it is impossible to tell how much of the water (and steam) of hot springs and fumaroles is of surface and how much is of magmatic origin.

While there are various types of hot springs, according to temperature or substances in solution, the most interesting are boiling springs and geysers. Warm carbonated springs that deposit travertine have been already described in Chapter VI.

Boiling Springs. — Actively boiling springs are a feature of many volcanic regions. Many of them occur in Yellowstone Park, especially in the different geyser basins (Fig. 173). They grade from pools that are hot but rarely boil, or else simmer quietly, into springs that boil strongly and steadily, and even some that boil more or less violently and with somewhat explosive energy, interrupted by short periods of repose. The latter form transitions to the geysers mentioned beyond. So long as the supply of water is sufficient to enable the spring to have an overflow it remains limpid, and it usually has a deep blue or green color; but if the evaporation through boiling is equal to the inflow, the water is more or less turbid from particles of disintegrated rock, and eventually becomes a mass of boiling mud. Such hot springs are called "paint pots" or "mud pots." In some of them the mud is white; in others it is tinted yellow, red, purplish, or black by oxides of iron and manganese. The mud as it increases in amount becomes so thick and viscid as to prevent regular ebullition, and, owing to the accumulating steam pressure, the paint pot boils spasmodically and with some violence, the mud being thrown into the air and about the vent, where it collects in considerable masses. These are known as *mud volcanoes*, or mud geysers. They usually mark a declining stage of activity in the life of a hot spring.

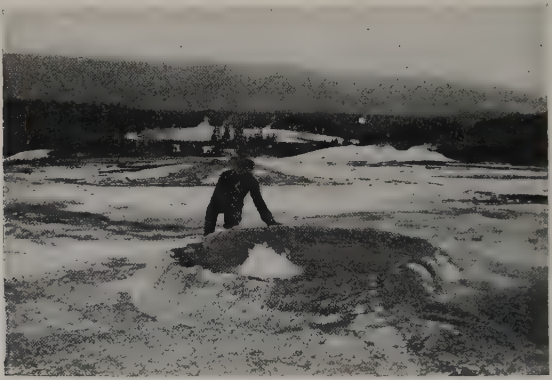

Fig. 173. — "The Devil's Punch Bowl." A hot spring boiling in the cup-like deposit of geyserite it has formed. The opening is several feet in diameter. Upper Basin, Yellowstone Park. (U. S. Geol. Surv.)

Geysers. — This term, from an Icelandic word meaning to gush, is applied to certain hot springs that at intervals spout a column of hot water and steam into the air. Depending on the size of the geyser and its special peculiarities, the height to which the column of water is ejected ranges from only a few feet up to several hundreds; the eruption may last a few minutes or several hours; the quantity of water discharged varies from a small quantity up to many thousands of gallons; the jet may play steadily and continuously straight up, or it may be fitful, be composed of minor jets, or be thrown in inclined directions. The interval between eruptions may be a definite number of minutes or hours, or it may be irregular, and several days may elapse between eruptions. Each geyser has in these ways its own peculiarities. As they are boiling springs of a special kind they are not common and are almost wholly confined to three principal regions: Yellowstone Park, Iceland, and New Zealand.

Some geysers consist at the surface of basins, ranging from several feet to a number of yards across, and rather deep. The sides and edges of the basins usually are beautifully ornamented by the deposits of silica described beyond, and they terminate at the bottom in tubes or fissures leading to the heated depths below, as shown in the diagram, Fig. 176. The tubes and basins are, except after eruptions, filled with water at or near the boiling point. In other types the geysers by their deposits have built up mounds, or cones, of silica, from a foot or two to several yards high, which form upward continuations of the pipes (Fig. 174). Of the Yellowstone geysers the most celebrated perhaps is the one known as " Old Faithful," which for many years after its discovery had a very regular interval between eruptions of about 65 minutes. It is now less regular, ranging from 60 to 80 minutes. (Fig. 175.)

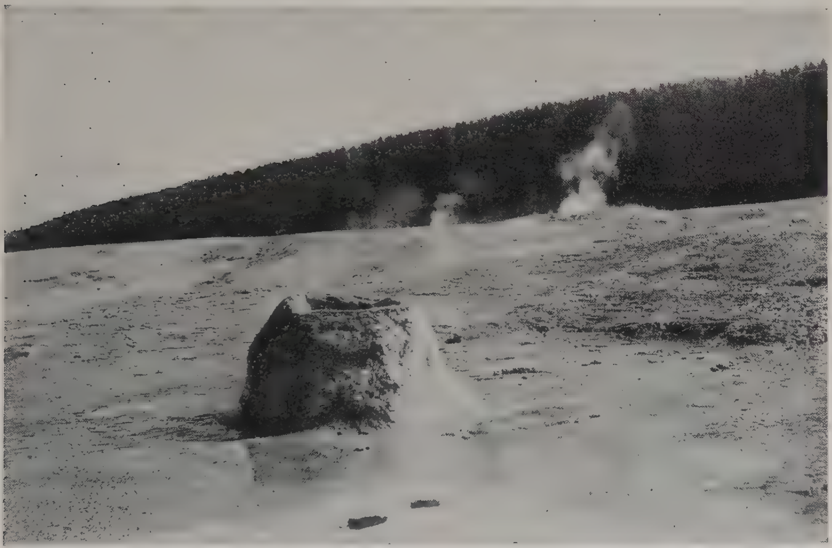

Fig. 174. — Silica cone of the Beehive Geyser, Upper Geyser Basin, Yellowstone Park. (U. S. Geol. Surv.)

Cause of Geyser Action. — The intermittent eruptive action of geysers depends on the relation between pressure and the boiling point of water, as was pointed out by Bunsen in connection with the great geyser in Iceland. The boiling point of water under the ordinary pressure of the atmosphere at sea level is 212° F.; increase of pressure raises it, a decrease lowers it. Thus the boiling point at the bottom of a column of water will be raised by the pressure of the superincumbent column above it; as shown in Fig. 176, it will gradually rise as we follow the tube from the surface downward. If, however, the cavity or fissure is large and open, the heated water below will rise, convection currents will be established, mixing the water, so that it will have nearly, though not quite, the same temperature in different parts of the cavity, and a regular boiling spring will result. But if the tube is long, narrow, tortuous, or constricted, convection will be prevented or restrained, and the water must boil in different levels at different temperatures corresponding to the pressures. Suppose at a level corresponding to 230° in the figure the boiling point is reached; bubbles of steam are formed, the column of water above is raised a little by the expansion, the bubbles of steam rise in the cooler liquid above and collapse, the column of water settles back with jarring, thudding sounds commonly heard before eruption. The temperature of the water will gradually rise until it is just about at the boiling point for each level corresponding to its depth and pressure. Finally when a sufficient volume of steam

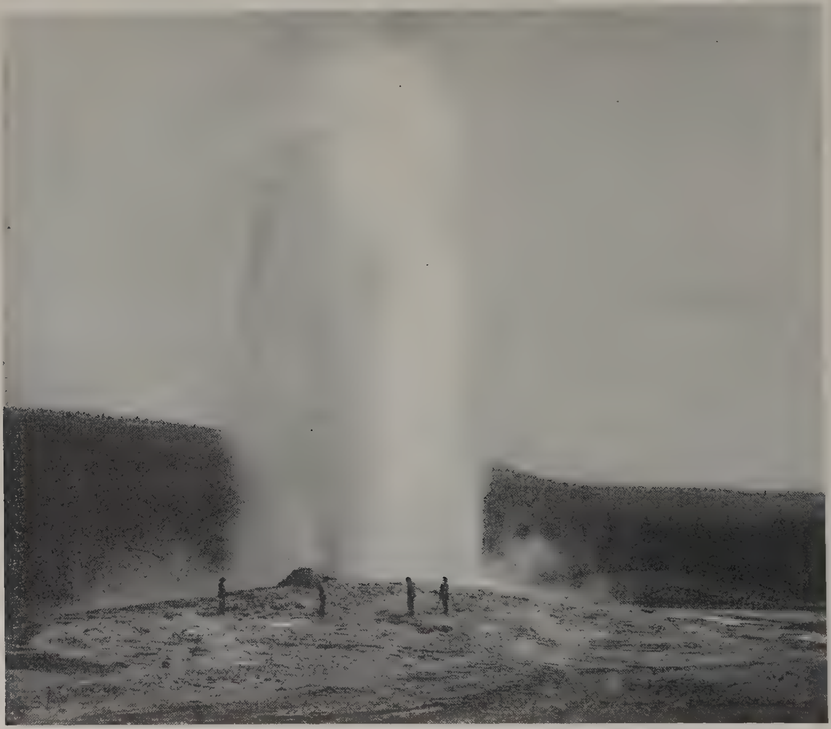

Fig. 175. — Old Faithful Geyser in eruption. Yellowstone Park. (U. S. Geol. Surv.)

is formed in the lower part of the geyser tube, the expansion will cause some of the water in the basin or cone at the top to overflow. This overflow lowers the pressure throughout the tube, and the water at each level, being now heated above the boiling point for the diminished pressure, will immediately flash into steam, and a mingled column of steam and hot water will be driven roaring out of the pipe into the air. After the eruption is over, the system fills again by inflow of ground water into the geyser tube, and the process is repeated.

The varied forms of fissures, underground conduits, and water supply account for the peculiarities shown by different geysers. It was found by accident that adding alkaline substances, such as soap or lye, to the waters of geysers causes some of them to erupt very quickly. The government had to put a rigorous ban on the " soaping " of the geysers of the Yellowstone National Park, in order to prevent ruining them.

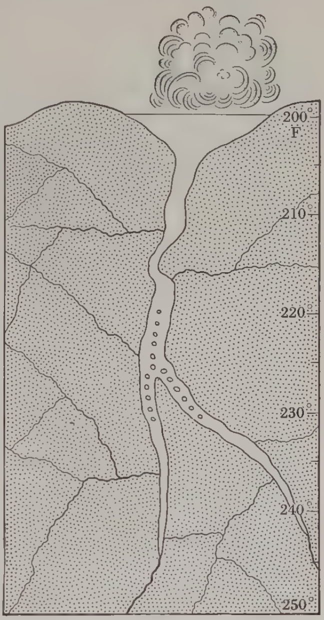

Fig. 176. — Vertical section to illustrate conditions necessary for geyser action.

Hot-Spring Deposits. — It has been previously shown that warm springs, especially if they contain carbon dioxide in notable quantity and come up through limestone beds, form deposits of calcareous tufa, or travertine. (See Chapter VI.) But the waters of boiling springs and geysers, which occur only in regions of recent volcanic activity, are mostly alkaline and carry *silica* (SiO_2) in solution, which they deposit as a whitish material. This deposit ranges from compact to spongy in texture, and is known as *geyserite*, or more commonly as *siliceous sinter*. It forms the geyser cones, or is deposited as incrustations, much of it of great beauty, in and about the margins of the hot-spring and geyser basins. The geyser waters are dilute, in fact, so dilute as to be tasteless, and the rate of deposition is very slow when it occurs only through evaporation but is hastened by the action of organisms. As in the case of travertine, the deposition of silica is largely due to its secretion by low forms of vegetable life (diatoms and algae, the latter related to seaweeds), which flourish in the warm waters and even in the hot waters. The beauty of many of the pools is greatly enhanced by the rich coloring that these growths give to them.

Besides silica, the hot springs deposit other substances. The waters of some springs are acid and deposit sulphur and alum salts; and from other springs sulphides of arsenic and of metals are deposited, thereby throwing light on the processes by which ore bodies are formed.

Economic Utilization of Fumarole Fields

In recent years it has become apparent that fumarole fields may be developed so as to yield large supplies of steam for power generation. The fumarole field in the volcanic area of Tuscany, north of Rome, was the first to be developed. Wells have been put down to depths of 600 feet, and the flow of steam, as well as the temperature of the steam, has been found to increase as depth is gained. About 30,000 horsepower is generated and is transmitted to Florence, 60 miles away, Pisa, and other cities.

At " The Geysers," in the Coast Ranges of California, 40 miles north of San Francisco, is an area of 35 acres containing a few feeble fumaroles and some small but very hot springs. The name " Geysers " is a misnomer, however, as none of the hot springs is periodically eruptive. In 1921 the idea was conceived of drilling wells in this area to develop a flow of steam for power purposes. So far eight wells have been put down, the deepest being 650 feet, and copious supplies of superheated steam have been developed. In fact, it is estimated that four of the wells will on the average deliver more than 1300 horsepower each. As in Tuscany the deeper a well is drilled the greater the steam flow and the hotter the steam.

CHAPTER XI

THE IGNEOUS ROCKS

The igneous rocks form one of the great divisions of the rocks that make up the crust of our planet. As their name implies, heat was an essential factor in their origin, and they may be defined as those rocks *which have been formed by the solidification of molten matter that originated within the Earth.* Such molten matter, as was explained in the discussion of volcanic action, is commonly called *magma,* a term we shall frequently use.

Distinguishing Characters. — The features that distinguish the igneous from the sedimentary and metamorphic rocks consist partly in the relation that the igneous masses exhibit towards other rocks with which they occur or with which they are in contact (a relation that we term their mode of occurrence), and partly in the characters that become evident when the rock itself, instead of the mass of which it is a part, is closely examined.

Igneous rocks do not of course contain fossils, nor do they show as a rule the parallel or banded appearance of the stratified rocks. They have certain distinctive peculiarities in the arrangement of their component mineral grains (the *texture,* as it is called). Some igneous rocks, indeed, are more or less made of glass, which at once betrays their origin, because glass is formed only by the chilling of molten material. The textures will be described when the different kinds of igneous rocks and their classification are considered; first we shall discuss the ways in which masses of igneous rocks occur as elements in the architecture of the Earth's crust.

OCCURRENCE OF IGNEOUS ROCKS

Intrusive and Extrusive Rocks. — There are two chief modes of occurrence of igneous rocks: the *intrusive* and the *extrusive.* In the intrusive mode of occurrence the magma, at the time it was rising from the depths, stopped before reaching the surface, and consequently it cooled and solidified under the cover of the rock masses of the Earth's outer shell. In the extrusive mode of occurrence, the magma attained the surface: it was extruded upon it, and has solidified there. The extrusive rocks are sometimes called *effusive* and sometimes *volcanic* rocks.

Although the division of igneous rocks into intrusive and extrusive is a natural one, the two classes are closely connected and, in fact, grade from one into the other. The magma of any extrusive mass came up through some passageway from below; this passageway remained filled, and eventually the magma in it solidified into rock. Consequently in theory every extrusive body that occurs on the Earth's surface is connected with an intrusive mass occurring below. In some places this connection of the extrusive mass with its root may be seen, but more generally the extrusive covers the root or has been separated from it by erosion, and the former continuity has been destroyed. It is clear also that we must think of the intrusive prolongation as extending downward and connecting with some greater mass of magma (or rock) below, of a nature to be presently described. See in this connection what has been said regarding the relation between volcanoes and deep-seated masses of magma in the preceding chapter.

Both intrusive and extrusive rocks have various modes of occurrence. The mode of occurrence of intrusive masses depends on how the intrusive mass is related to the rocks that enclose it; and the mode of occurrence of extrusive masses depends on the conditions under which the magma was ejected. We shall begin with the modes of occurrence of intrusive masses; but it should first be recalled that, inasmuch as these intrusive masses were covered at the time of their formation by the rocks into which they were intruded, they can be exposed at the Earth's surface and thus laid open to observation only after erosion has carried away the cover and has disclosed their intrusive character. In some places, where the magmas were intruded under a thin cover, the time necessary to do this may have been short; but in other places, where the igneous masses were deeply buried, it may have been exceedingly long.

Intrustive Modes of Occurrence

The principal kinds of intrusive igneous bodies are *dikes*, *sills*, *laccoliths*, *necks*, *stocks*, and *batholiths*. Several other modes of occurrence are recognized, but as they have not the importance of those mentioned, they will be treated for simplicity's sake as modifications of them. The simplest form of intrusive body is the dike and this will be considered first.

Dikes. — A dike results from the solidification of magma that has filled a fissure in preëxistent rocks. Consequently, it is a tabular mass: its length and breadth are great compared with its thickness. It may "cut," that is, pass through, rocks of any kind — igneous, sedimentary, or metamorphic. In the sedimentary rocks it must by definition cut the planes of stratification at an angle (Fig. 177); if the igneous mass lies parallel to the bedding planes it is termed a sill. A dike may be a few yards or many miles long; it may be a fraction of an inch, or many hundreds, or even some thousands of feet thick. Most dikes are from 2 or 3 up to 20 feet thick; the length varies greatly. A great dike in the north of England extends for over 100 miles.

Some dikes have attained the Earth's surface and given rise to outflowings of lava, but others have not reached it and have become exposed only by subsequent erosion. Some dikes were the canals that fed larger intrusive bodies above them, such as the sills and laccoliths to be next described. In the process of erosion, a dike may be more resistant than the surrounding rocks and hence is left projecting as a wall; some, however, are less resistant and form ditches; from these features the name is derived, especially from its resemblance to the more prominent wall, for dike means both wall and ditch. The rock of some dikes is divided into blocks by joints, and very commonly the blocks are columns lying perpendicular to the walls of the dike, like

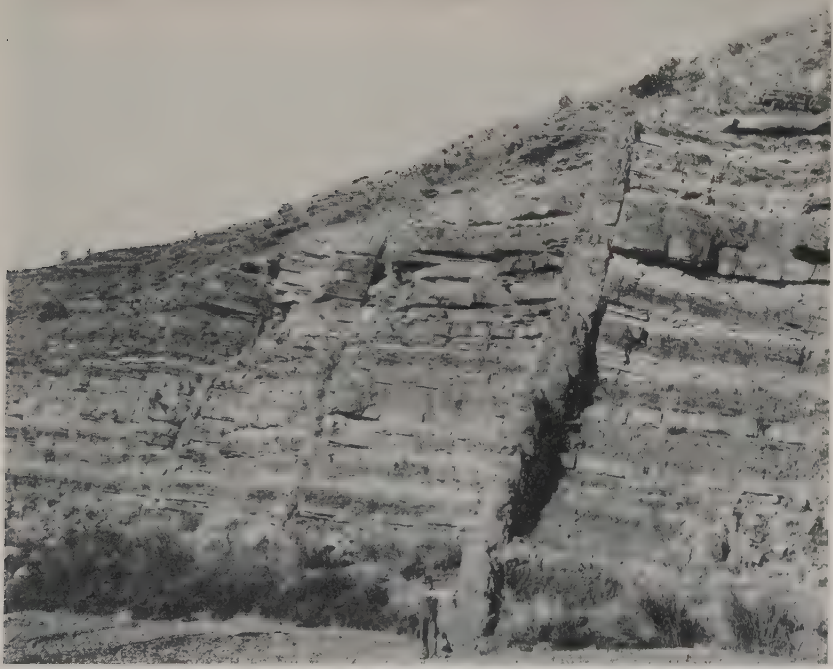

Fig. 177. — Dikes cutting across sedimentary beds. Alamillo Creek, N. Mex. (U. S. Geol. Surv.)

a pile of cordwood, an arrangement whose origin is described later under columnar structure. Dikes occur at many places in more or less well-defined systems, and around volcanic centers are likely to be radially disposed.

Sills. — It is not uncommon to find, where magma has been intruded in bedded rocks, that it was injected as layers between the beds. An igneous mass that thus lies concordantly between the bedding planes is known as a *sill,* or an intrusive sheet. Sills range in thickness from a foot or so to several hundred feet, and some of them are many square miles in area.

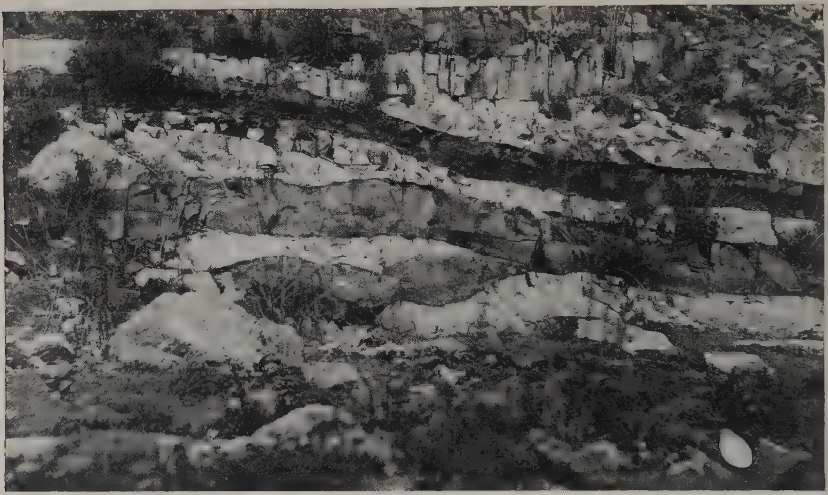

Fig. 178. — Sills of igneous rock. Cottonwood Canyon, New Mexico. They follow the bedding planes generally, but break across the beds locally, at a low angle. (U. S. Geol. Surv.)

Some sills break dike-like across the strata and then continue along a new horizon. Sills may be distinguished from flows of lava that have been buried by deposits of later sediments by the fact that the rock composing the sills is of the same compact nature at top and bottom, i.e., is not slaggy or scoriaceous like a lava flow; and by baking and alteration of the overlying sediments from contact with the intrusion while it was hot. The surface of a lava flow is usually spongy, ropy, and slaggy, and a flow could of course exert no action on beds not yet deposited upon it. Sills are most likely to occur where larger intrusions of magmas, such as laccoliths and stocks, have taken place, as accompanying features in the surrounding strata. Prominent and scenic land features are produced by erosion in regions where thick sills occur and the strata have been dislocated and upturned. This is illustrated in the Palisades of New York and New Jersey, and in some of the trap ridges of southern New England.

Laccoliths. — The laccolith in its typical development is a lenticular or dome-shaped mass of igneous rock that was intruded into sedimentary rocks between the bedding planes. It has a flat floor, and is more or less circular in ground plan. If during the formation of a sill, magma is supplied more rapidly from below than can easily spread laterally away from the supply channel, the overlying strata will be arched up, as if by a hydraulic press, and a thick lens of liquid rock will be produced, giving rise on solidification to a laccolith. Such masses range from a few hundreds of feet to more than a mile thick at the center, and a few hundreds of yards to many miles in diameter. An ideal section of a laccolith is shown in Fig. 179, and a photograph of one from which the cover has been removed by erosion, exposing the igneous rock, is seen in Fig. 180.

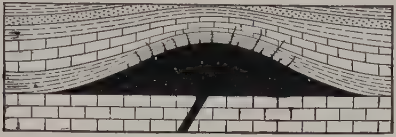

Fig. 179. — Section of a laccolith. The black area is the igneous rock.

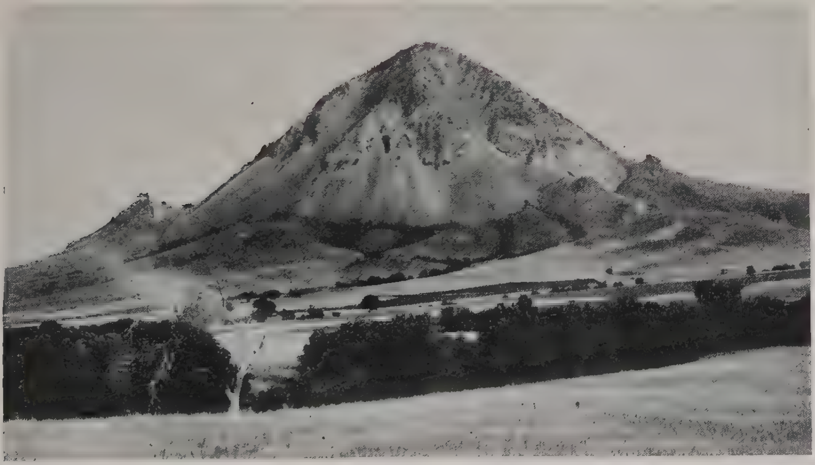

Fig. 180. — Bear Butte, a laccolith denuded of its cover and the igneous mass laid bare. The ring of upturned eroded strata is seen about its base. Black Hills, South Dakota. (U. S. Geol. Surv.)

Although the above statement gives the idea of a typical laccolith, many departures from this arrangement are found in the actual occurrences. Some laccoliths are oval or quite irregular in ground plan, instead of circular; and in section some have the shape of a wedge instead of the symmetrical form shown in Fig. 179. According to their degree of flatness, all transitions into sills occur. Some thin out into sills, or are accompanied by sills on the flanks of the arches, and thus are compound in structure.

Laccoliths, more or less exposed by erosion, are conspicuous features in many parts of western North America, where they were first discovered by Gilbert in the Henry Mountains of southern Utah. Subsequently they have been found in various parts of the world and are therefore a not uncommon form of intrusion. Moreover, some of the more recently recognized laccoliths are of immensely greater size than the classic laccoliths of the Henry Mountains, the largest of which has a volume of 10 cubic miles. The great Duluth laccolith on the northwest shore of Lake Superior is estimated to represent the injection of 50,000 cubic miles of magma.

Necks. — When a volcano becomes extinct, the column of magma that filled the conduit leading to unknown depths below solidifies and forms a cylindrical mass of igneous rock. Erosion will in time cut away a great part of the ashes and lavas of the cone, leaving this more compact and resistant rock projecting, as shown by the line *abc* in Fig. 169. If the level of erosion eventually descends into the rocks that form the basement on which the volcano stands, and all of the ashes and lavas are thus swept away, only the projecting mass remains to mark the former site of the volcano. Such a mass of rock is known as a *volcanic*

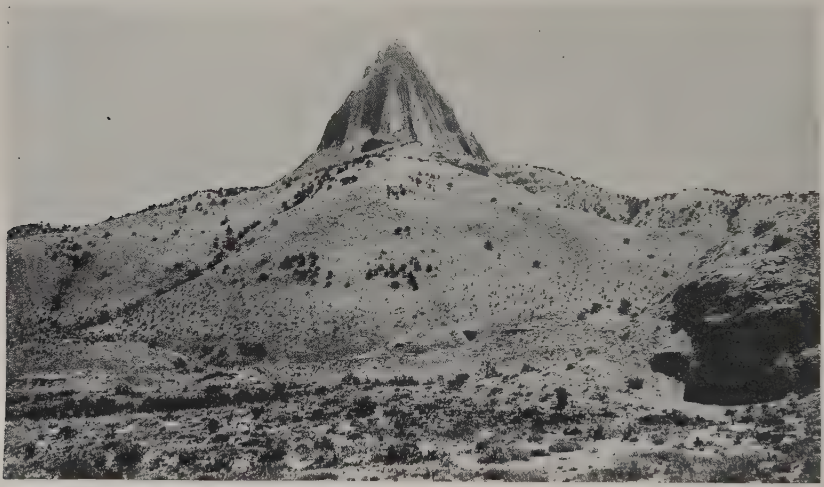

Fig. 181. — Alesna volcanic neck, Mt. Taylor region, New Mexico. (U. S. Geol. Surv.)

neck (Fig. 181). It is commonly more or less circular in ground plan and from a few hundred yards up to a mile or more in diameter. The rocks about volcanic necks are likely to be cut by a radiating system of dikes, and commonly, if stratified, injected with sills. The significance of volcanic necks has been previously explained in the discussion of volcanoes.

Stocks. — This term is applied to certain large domal bodies of intrusive rock that in the form of magma have ascended into the upper levels of the Earth's crust and there solidified. They have become visible because erosion has stripped off the covering rocks. They have as a rule a more or less circular or oval ground plan. Their outer surfaces cut across the inclosing rocks, are more or less irregular, and the masses usually widen in extent as they descend. Their size ranges from a few hundred yards to several miles in diameter. As they are likely to form protuberant topographic features after being exposed by erosion, they are sometimes called *bosses*. The distinction from a volcanic neck is not one of size alone, though necks tend to be smaller than stocks, but in that the term " neck " is employed only when there is evidence that the igneous body has functioned as the supply conduit of a volcano. Some stocks were doubtless necks, but this cannot now be proved.

Batholiths. — A batholith is a huge intrusive mass of igneous rock. According to some definitions, a batholith is " bottomless," that is, it extends indefinitely downward into the Earth's crust, in contrast to a laccolith, which rests on a floor. The " bottomlessness " is of course pure hypothesis. A batholith differs from a stock only in its much greater size, as some are exposed over many thousands of square miles of surface. Arbitrarily, an intrusive igneous body less than 40 square miles (100 square kilometers) in areal extent is called a stock; if larger than 40 square miles, a batholith. Some stocks, as shown in Fig. 182,

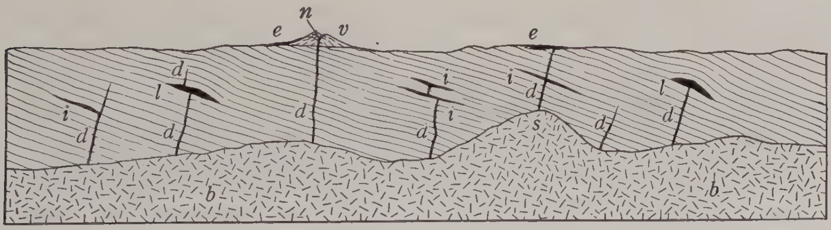

Fig. 182. — To illustrate the occurrence of igneous rocks: *b*, batholith intruded into tilted sedimentary rocks; *s*, stock; *l,l*, laccoliths; *i,i*, sills; *e,e*, lava flows; *d,d*, dikes; *v*, volcanic cone built of tuffs and breccias; *n*, filling of the volcanic conduit, which probably will become a neck through erosion. Horizontal distance shown, twenty miles; vertical distance, three miles.

are merely dome-like protuberances from the body of an underlying batholith; they have been aptly called cupolas.

The largest batholith in the United States is the Idaho batholith in central Idaho, extending over an area of 16,000 square miles. The causes for the rise of so stupendous a mass of molten rock matter into the higher levels of the Earth's crust and the processes by which it makes room for itself are among the most fundamental and fascinating problems of Geology.

Extrusive Igneous Rocks

The modes of occurrences of the extrusive igneous rocks have already been described in connection with volcanoes and extrusions of lava, and what has there been said in regard to them may be profitably consulted in this connection. Lava flows and beds of volcanic tuff and breccia are widely distributed. They occur wherever volcanic activity is being or has been displayed, and their presence is, indeed, one of the surest indications of former volcanism in places where it has long since died out. We are thus able to recognize that volcanoes formerly existed in various parts of the eastern United States and Canada, from Nova Scotia to Georgia. These volcanic rocks occur in vast quantities piled up in places thousands of feet in thickness in the Rocky Mountains, where, as in western Wyoming, serried mountain peaks have been sculptured from them by erosion.

Age of Igneous Rocks

The geologic period when a given mass of igneous rock was erupted or intruded is determined by ascertaining its relations to the rocks with which it has come in contact. Thus, if a body of igneous rock, such as a dike, cuts through another body of rock, it is manifestly the younger of the two. If it cuts across stratified beds it is younger than they are, and lavas of course are more recent than the rocks upon which they lie. If a sheet of igneous rock lying concordantly between strata has affected the beds both above and below it (see contact metamorphism, page 307), it is younger than both. If it has not baked or otherwise metamorphosed the overlying beds, the sheet may be a lava flow and older than they are. It is thus usually easy to tell when an igneous mass is younger than other rocks by examining its contacts with them. The age of the stratified rocks is determined by the fossils that occur in them, and this determination gives a basis for working out the geologic date of igneous masses with which the fossiliferous rocks are in contact.

KINDS AND CLASSIFICATION OF IGNEOUS ROCKS

Introductory. — The features by which the igneous rocks are distinguished have been already mentioned in a preliminary way in discussing the products of volcanoes, but they should now receive the attention that they demand. Igneous rocks are divided into different kinds on the basis of two properties: first, their texture; and second, their composition. Each of these properties requires explanation.

Texture. — The most obvious thing about an igneous rock is its texture. By *texture* is meant the size of the component grains and the shape and arrangement of the grains. Thus if the grains are as large as peas, we say that such a rock is *coarse-grained* in texture; if the grains are the size of those in granulated sugar, we say that the rock is *fine-grained;* whereas, if the particles are so minute that they cannot be discriminated by the unaided eye and the rock looks as if it were a homogeneous substance, we say that it is *aphanitic*[1] in texture.

The texture depends on the *rate at which the magma cooled.* For, if the magma is too hot, as previously explained, crystallization cannot take place, and no crystals will begin to form until the temperature has fallen far enough; then they will begin to separate from the magma and, if the cooling is very slow, they will have time to grow to large size, thus producing a coarse-grained rock. But, if the cooling is rapid, more and more new centers of crystallization will be forced to form, and, if the process is thus hurried, instead of few crystals growing to large sizes, the rock will consist of a large number of smaller particles and will, therefore, be fine-grained in texture. And with still more rapid cooling the particles may be so minute that they are not discriminable by the unaided eye and the resultant rock is of aphanitic texture. Analogy will now carry us one step more: we can conceive that the cooling may take place with such great rapidity that the magma will solidify into a homogeneous substance before any crystallization, which consists in the molecules arranging themselves together to form definite solid compounds, can occur. In this event the result will be a *glass*, or a glassy rather than a stony texture, a case that is by no means uncommon.

To sum up, then, we see that igneous rocks may be *coarse-grained*, *fine-grained*, *aphanitic*, or *glassy* in texture, and that which of these textures is developed depends on the rate of cooling of the magma.

Porphyry: Porphyritic Texture. — In what has been said so far regarding the texture of igneous rocks, it has been tacitly assumed that the component mineral grains in any given rock are of uniform size, or that the rock is evenly *granular*, as it is called. Not all igneous rocks, however, are evenly granular. Inspection shows that many of them are composed of crystals of two sizes: some crystals that are larger and more distinct and which are embedded in a matrix of much finer grains. An igneous rock having this texture is called a *porphyry* (Fig. 183). The matrix of a porphyry is termed the *groundmass*, and the large crystals embedded in the groundmass are called the *phenocrysts* (clearly discernible crystals). Porphyritic rocks are common.

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[1] *Dense* is often incorrectly used in America as a synonym for aphanitic. It is also used correctly to mean "of high density" and this double usage leads to ambiguity.

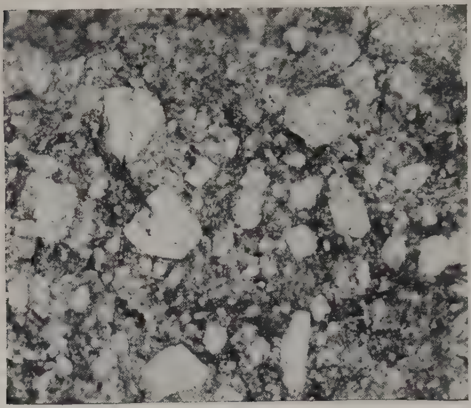


Fig. 183. — An igneous rock with porphyritic texture.

The groundmass itself varies widely in grain size in different porphyries: it may be medium-grained, fine grained, aphanitic, or glassy; most commonly it is aphanitic, as in the extrusive rocks. The phenocrysts also vary widely: they may be as large as walnuts or as small as grains of sand; they may be abundant or comparatively few. But in all porphyries there is this *contrast* between *sizes* of crystals, between groundmass and phenocrysts, which makes the essence of a porphyry. The student should guard against thinking that the porphyritic texture is a contrast of *colors;* thus a rock consisting of grains of light-colored quartz and feldspar, in which are embedded a few black crystals of mica, all grains being of about the same size, is *not* a porphyry.

Relation of Texture to Geologic Mode of Occurrence. — Obviously an *intrusive* mass of magma that is surrounded and blanketed above by other, older rock masses must lose heat much more slowly than an *extrusive* one, which is poured out on the surface in the form of lava. Hence, as a coarse-grained texture is the result of slow cooling, we naturally associate it with intrusive masses, and, conversely, we regard the aphanitic and glassy textures as belonging to the products of extrusion — the lava flows. But it is also clear that the size of the mass will greatly influence the rate of cooling, since a very large mass cools more slowly than a small one. Thus, the rocks that make up great batholiths are coarse in texture, whereas the rocks of dikes and sills tend to be much finer-grained. On the other hand, the central portion of an extremely thick lava flow may cool with sufficient slowness to develop a medium-grained texture, whereas a magma that was forced into a narrow fissure in cold rocks might be chilled so quickly as to assume an aphanitic, or even glassy texture. Thus various modifications of the general rule according to particular cases can be easily imagined; nevertheless this general rule, that the intrusive rocks are medium- to coarse-grained and the extrusive rocks are fine-grained to aphanitic, holds true.

Although the rate of cooling is the most important factor that determines the texture of igneous rocks, as discussed above, it is not the only one. The subject is too complex for detailed treatment in this work, but it may be mentioned that the chemical composition also has its influence. Under similar conditions of cooling, basic magmas (those low in silica and high in iron and magnesia) tend to assume a coarser grain than those composed of much silica, alumina, and alkalies. The reason is that the basic magmas are more fluid than the siliceous magmas, as already explained, and thus when they crystallize this mobility of the molecules permits the crystal grains to grow to larger sizes.

Also the presence of the included gases that magmas contain, especially the water, increases the fluidity and thereby promotes a coarser crystallization. This is very notably shown in and around certain intrusive masses by dikes that are made up of large and even huge crystals of quartz, feldspar, and mica. Crystals several feet in diameter are not uncommon. The very coarse masses of this composition are known as *granite pegmatites*, and from them are obtained the plates of mica that are used commercially. In allusion to their giant grain the pegmatites are sometimes called giant granites.

Composition of Igneous Rocks. — Since igneous rocks are produced by the solidification of magmas, their composition will obviously depend on the chemical composition of these fluids. It has been already shown (Chapter X) that a magma consists of two parts: a volatile part, consisting of the fugitive constituents — water vapor, carbon dioxide, sulphur fumes — amounting to a few per cent, and a non-volatile part, consisting of the fixed constituents, chiefly molten silicates. Although for reasons we cannot now consider, the gaseous part is important in rock formation, it is essentially the molten silicates that give rise to the rocks and are the chief constituents of magmas. While it is evident that a magma cannot be analyzed directly, still the fixed constituents and their relative proportions can be ascertained by analysis of the cold and solid rock. Many thousands of igneous rocks from all parts of the world have been analyzed and the following results have been obtained, which show the ranges in the amounts of the various constituents.

Silica, always present; ranges from 35 to 80 per cent.
Alumina, ranges from 0 to 25 per cent.
Oxides of iron, 0 to 20 per cent.
Magnesia, 0 to 45 per cent.
Lime, 0 to 20 per cent.
Soda, 0 to 16 per cent.
Potash, 0 to 12 per cent.

It must not be concluded from the above table that any and all sorts of mixtures of these oxides can occur within the limits shown. Certain general laws govern their associations. It is generally true that large percentages of potash and soda (alkalies) in a rock are accompanied by a correspondingly large amount of silica and consequently by small amounts of the other three metallic oxides. Conversely, large percentages of magnesia, lime, and iron oxides are likely to be associated, and these go with low silica, the alkalies being small or wanting. These reciprocal relations are of fundamental importance in igneous rocks, and it will be recalled that they have been pointed out before, because the nature of volcanic activity and the kinds of volcanoes in large measure depend on them.

Crystallization. — It is a familiar experiment that if a liquid containing a salt, zinc sulphate for example, is boiled down and concentrated to a certain point, all of the zinc sulphate can no longer remain in solution, but will begin to appear as a solid in the form of crystals. If the hot solution is allowed to cool, more crystals of the salt will be formed, since hot solutions as a rule can contain more salt than cold ones. In analogy with this, a magma can be regarded as a solution of various minerals. If the magma cools with sufficient slowness, the dissolved matter in it will separate from it as crystals. This crystallization will proceed as the temperature falls until the whole magma has turned into a mass of solid crystal grains. The molten liquid has become *stone.* The minerals separate from any given magma in a definite order, which is governed by their solubility in that magma.

Kinds of Minerals. — The more important minerals that form the igneous rocks are the following:

Feldspar Group	*Ferromagnesian Group*
Orthoclase Feldspar	Mica (Biotite)
Plagioclase Feldspar	Pyroxene
	Hornblende
	Olivine
Quartz	Magnetite (Iron Ore)

Of these minerals, *feldspars*, *quartz*, *pyroxene*, *hornblende*, and *biotite* are the most important in forming igneous rocks, and consequently the student should make careful note of them. For details regarding them the Appendix that deals with the minerals mentioned in this work should be consulted.

Furthermore, since it was shown that the chemical composition of magmas varies, it is evident that the relative quantities of the minerals in the resulting rocks will also vary. Thus, a siliceous magma, in which soda, potash, alumina, and silica are the chief substances, will form a rock that consists mostly of feldspars, whereas a basic magma, in which lime, magnesia, and iron oxides are high, will make a rock that contains mostly pyroxene, hornblende, and other *ferromagnesian* minerals, as they are called in allusion to the iron and magnesium in them.

Classification of Igneous Rocks

The features by which the igneous rocks may be classified have now been explained. We have seen that the igneous rocks vary in texture and in composition. Both of these variables will be used as factors in building up a classification of the igneous rocks. By employing texture as the principal criterion we at once obtain five major classes: I, even-granular, in which all the minerals are of about the same size and are sufficiently large to be identified by the eye alone or aided by a pocket lens; II, porphyritic-granular, in which certain minerals in virtue of their large size contrast conspicuously with those which surround them, thus forming a porphyry having an even-granular groundmass; III, porphyritic-aphanitic, in which the conspicuous crystals — the phenocrysts — are set in an aphanitic groundmass; IV, aphanitic, in which none of the constituents are distinguishable; and V, glassy, in which few or none of the constituents have crystallized. These five classes are termed *massive* rocks, and for the sake of completeness Class VI is added to take care of the fragmental products of volcanic eruptions. This order from Class I to Class V marks in a general way the decreasing amount of easily recognizable minerals in rocks: in Class I, all the constituents are easily recognizable by the unaided eye; in II, the phenocrysts are easily distinguishable, the constituents of the groundmass less readily; in III, the phenocrysts alone are distinguishable; and in IV, none of the constituents can be recognized.

Each of the major classes is then subdivided on the basis of composition — on the kinds of minerals present and the proportions in which these minerals occur. It is to these subdivisions that the actual rock names are given. For example, an even-granular rock that is composed of feldspar, quartz, and a dark mineral, generally biotite, is called granite.

By applying these principles, the following classification of igneous rocks is obtained, as shown in the subjoined table.

TABLE OF IGNEOUS ROCKS

	Major Classes (based on texture)	Subdivision of Major Classes (based on mineral composition)			
		Light-colored minerals, chiefly feldspar, predominate		Dark minerals predominate	Dark minerals entirely
I	Granular (with grains interlocking)	Granite (has quartz)	Diorite (has no quartz)	Gabbro, Dolerite (grain size is intermediate between that of gabbro and basalt)	Peridotite, Hornblendite, Pyroxenite
II	Granular (as above) and porphyritic	Granite Porphyry (has quartz)	Diorite Porphyry (has no quartz)	Gabbro Porphyry	
III	Porphyritic, with aphanitic groundmass	Rhyolite (contains phenocrysts of quartz)	Andesite	Basalt	
IV	Nonporphyritic, aphanitic	Felsite			
V	Glassy	Obsidian Pitchstone Pumice		Basalt Glass	
VI	Fragmental	Volcanic tuff and breccia			

Note: Syenite is briefly mentioned in the text.

Remarks on the Table. — Leaving out of account the glassy rocks, which are rare, and the tuff and breccia, which are described on page 222, the following remarks may prove of service in understanding the classification of igneous rocks shown in the table.

All rocks in the same horizontal column have the same texture.

All rocks in the same vertical column are essentially of the same chemical composition; for example, granite, granite porphyry, and rhyolite are alike in chemical composition. In physical appearance, however, they differ notably: a granite differs somewhat from a granite porphyry, and vastly from a rhyolite. These differences, as already pointed out, are mainly the results of the different rates of cooling — the granite has cooled extremely slowly, whereas the rhyolite has chilled rapidly.

The rocks in which the light-colored minerals predominate are light in color and light in weight, i.e., they are of low specific gravity. The rocks in which the dark (ferromagnesian) minerals predominate are dark in color and heavy in weight. The range in specific gravity — from 2.67 for the average granite to 3 for gabbro — is not large, but is sufficient after experience to serve as an aid to identification.

Although in the table each rock has been put in a separate compartment, in nature no rock variety is sharply bounded from its neighbors that are shown in the table. There are for example transitional varieties between granite and diorite and between granite and granite porphyry. No hard and fast boundaries set off any of the so-called rock species. These facts often make it difficult to classify a given rock. It may as well be recognized at the outset that precise determination of rocks is at best a difficult matter, and that to classify accurately with the unaided eye the finer-grained and especially the aphanitic rocks, is as a rule impossible. When the accurate identification of a rock becomes a matter of high importance, recourse must be made to the microscope.

Granite. — As may be seen from the scheme of classification, granite is composed chiefly of quartz and feldspar (of the variety orthoclase). It contains also as a rule a variable amount of flakes of mica, less commonly of hornblende, or both. These component minerals are roughly of the same size, and hence granite is said to be even-granular, or equigranular. As the quartz was the last mineral to separate from the magma, it is molded around the earlier minerals and occupies the angular interspaces between them. This habit of the quartz produces an intimately interpenetrating and interlocking arrangement. This interlocking, even-granular texture is so characteristic of granites that it is often called for short the granitic texture. It serves to distinguish rocks of Class I from all others. The average granite contains 60 per cent of feldspar, 30 per cent of quartz, and 10 per cent of dark minerals. There are many varieties of granite, based on color, texture, etc.

Granite is the most important intrusive igneous rock, and appears to be the main constituent of the foundation of the continental masses. As granite is formed at some depth in the crust, it is exposed at the surface only after prolonged denudation; hence it is seen chiefly in those parts of the continents bared by erosion — that is to say, in mountains or in regions so deeply eroded that the roots of former mountains are visible.

Syenite. — This is like granite in texture and composition but differs in containing little or no quartz. The syenites are not common rocks, nor as a rule do they occur in very large masses compared with granite.

Diorite. — Diorite is a granular igneous rock composed of feldspar and one or more dark minerals, in which the feldspar is more abundant than the dark minerals. The feldspar is mainly plagioclase, but it is generally difficult to recognize this fact with the unaided eye. The dark minerals are biotite, hornblende, or pyroxene, occurring either singly or together.

Gabbro. — Gabbro differs from diorite in that the feldspar is subordinate and the dark minerals predominate. Hornblende, pyroxene, and olivine are the common dark minerals; they occur singly or together; biotite, though present in some gabbros, is distinctly uncommon. Because of the prevalence of dark minerals, gabbros are dark and of high specific gravity. *Dolerite* is a convenient term for the basic rocks that are intermediate in grain size between basalts and gabbros.

Diorites and gabbros, while abundant as intrusive masses, do not commonly occur in extensive batholiths as the granites do. They are more common as smaller stocks, sills or dikes.

Peridotite. — Peridotite is generally composed of a mixture of ferromagnesian minerals, with olivine predominating. It is not common and usually occurs as minor intrusions, — dikes, sills, or small stocks. It is very interesting and important, however, as being the source of ores of chromium, nickel, and platinum, and of the diamond. It is generally very dark to black and heavy from the large amount of iron-bearing minerals present. The diamonds of South Africa occur in volcanic pipes composed of this rock, and they have been also found in similar intrusions of it in Arkansas.

Pyroxenite, as its name implies, is composed wholly of pyroxene, and *hornblendite* consists entirely of hornblende. They form as a rule bodies of small size; nevertheless, in places, as at the remarkable platinum deposits recently discovered in South Africa, pyroxenite occurs in vast volume.

Granite Porphyry, Diorite Porphyry, etc. — As may be seen by reference to the table of classification, there are various kinds of porphyry, depending on the coarseness of the groundmass and its composition, and on the kinds of minerals that are embedded in it as phenocrysts. Thus if the groundmass is as coarse as it is in granite we may have granite porphyry, or diorite porphyry. Feldspars are the most common phenocrysts; quartz occurs along with the feldspar phenocrysts, chiefly in granite porphyry; and dark to black flakes or prisms of mica, hornblende, or pyroxene occur in many porphyries. The porphyries are a very common class of rocks, occurring chiefly as minor intrusions: in dikes, sills, and laccoliths, and often in necks; they do not occur as batholiths. They also compose many lava flows. Intrusions of porphyry in the Rocky Mountain region are very common, and in many places are accompanied by valuable deposits of gold, silver, copper, lead, and other ores. Examples of this are seen at Leadville and other places in Colorado, Montana, Nevada, etc. A porphyry is shown in Fig. 183.

Rhyolite. — Rhyolite represents the aphanitic lava form of the magma that at depth consolidates as intrusive granite. It contains phenocrysts of feldspar, quartz, and biotite, and rarely of hornblende. The number of these phenocrysts varies within the widest bounds, so that there is every transition between nonporphyritic and highly porphyritic rhyolite. When the amount of the phenocrysts exceeds 25 per cent of the volume, the rock is by some called a rhyolite porphyry. The colors range from white to gray, pink, red, and purple. Rhyolites or andesites with inconspicuous phenocrysts or with sparse or no phenocrysts are termed *felsite.*

Andesite. — Andesites are of many colors, but in general they are darker than the rhyolites; dark gray is common. On the one hand they are transitional into rhyolites; on the other, into basalt. The average or typical andesite occupies the intermediate position. The darker andesites are of basaltic appearance, but unlike basalts their freshly broken thin edges are translucent when held in bright light. The phenocrysts in andesites commonly consist of striated feldspar and one or more dark minerals (hornblende, pyroxene, or biotite). Quartz phenocrysts are absent (distinction from rhyolite). Andesites that are crowded with prominent phenocrysts are by some called andesite porphyries.

Andesite and andesite porphyry are enormously abundant among the extrusive rocks of the globe. They are the chief products of the volcanoes that form the " circle of fire " surrounding the Pacific Ocean. In fact, it was because of their prevalence in the *Andes* of South America that they were given their name.

Felsite. — The difficulty and often impossibility of discriminating between rhyolites and andesites that are devoid of phenocrysts makes it necessary to use an elastic non-committal name. For the light-colored rocks of this class, namely those which are white, light to medium gray, light-pink to dark red, pale yellow or brown, purple or light-green, in short those that are not dark green, dark gray, dark brown, or black, the term felsite is often convenient.

Basalt. — When the color of the lava is very dark gray, dark green, brown, or black the rock is *basalt.* It is either compact or vesicular.

If the vesicles have become filled with some mineral, such as calcite, chlorite, or quartz, the fillings are called *amygdules* and the rock is termed an *amygdaloidal* basalt. Many basalts are without phenocrysts, but others contain numerous conspicuous phenocrysts, consisting of feldspar, olivine, or augite, or some combination of these. The phenocrysts are hard and have straight, clean-cut boundaries, whereas amygdules are generally soft and have irregular, roundish, or elliptical boundaries. The effusive occurrence of basalt has been already treated under volcanoes and eruptions. The enormous tracts of land in western America, in India, and elsewhere that were flooded by outflows of basalt have there been mentioned.

Dolerite is the name given to the coarser-grained basalts, in which the grains are so well developed that the constituent minerals are nearly or quite recognizable. There is no hard and fast line between basalt and dolerite on the one hand and dolerite and gabbro on the other.

Glassy Rocks. — Volcanic glasses occur as thin crusts on the surface of lava flows or where a lava flow has been very quickly cooled, and they are mostly limited to siliceous magmas. Brilliantly lustrous volcanic glass is called *obsidian,* and the duller and more pitchy variety is called *pitchstone. Pumice* is frothed glass. Obsidian was much used in past times by primitive peoples in making weapons, implements, etc.

Obsidians are commonly dark-colored or even black; and yet many of them correspond in chemical composition to rhyolite and granite. Hence they appear to contradict the general rule that nearly all rocks with this composition are light-colored. However, if a piece of black obsidian is chipped to a thin edge it transmits the light and loses much of its dark appearance. The deep coloring is the result of uniform distribution of a relatively small amount of dark material in the glass.

Basalt glass is of rare occurrence. Its formation requires extremely rapid chilling of basaltic magma.

CHAPTER XII

WARPING, FOLDING, AND FRACTURING IN THE EARTH'S CRUST

The outer shell of the Earth is not fixed and rigid, but undergoes changes that result in movement of some parts with relation to others. Evidence is overwhelming that this has occurred repeatedly in the past in all places where it is possible to examine the structure of the Earth's crust. Movements of the different parts of the outer shell have been not only up and down, but also back and forth in directions parallel to the Earth's circumference. From historic records it is known that movements have taken place, some abrupt and catastrophic; others so gradual that their results are perceptible only after a long time interval. Back of human history we read the record of crustal movements in obvious deformation of the rocks, ranging from broad, gentle bending or warping to more localized severe folding or fracturing.

All movements of the lithosphere, resulting in relative vertical or horizontal changes of position and in deformation of rocks, are comprehended under the general term *diastrophism.* Movements that affect all or a large part of a continent are termed *epeirogenic,* from the Greek *epeiros,* a continent. More localized disturbances related to mountain building are designated as *orogenic,* from the Greek *oros,* a mountain. These terms are useful in discussion; but a systematic treatment of deformation may well begin with the more obvious effects, of whatever kind, produced either in historic or in late geologic time, and proceed to the results of more ancient movements.

DEFORMATION IN LATE GEOLOGIC OR RECENT TIME

Datum Surface. — The fact and the amount of any recent movement are determined from the relative positions of features on the Earth. Most of these movements are in the vertical direction; and in order to determine the extent and the rate of change it is necessary to have a convenient horizontal surface to which reference can be made. The average level of the sea is the most logical surface for this purpose, as the shoreline, at mean tide level, is essentially horizontal throughout its whole extent. Any local or differential movement, upward or downward, of land bordering the sea is clearly evident to one who observes the peculiarities of shorelines.

Elevation. — The most striking proofs of uplift of the land consist in the locally elevated position of features that we definitely associate with the sea or its edge. Thus in various parts of the world outcrops of rocks with attached shells or skeletons of dead marine organisms, such as barnacles and corals, are found high above sea level. In some localities the rocks are pierced by tubes that were drilled and lined by a peculiar rock-boring marine animal (*Lithodomus*). Excellent evidence of continuous or recurrent elevation is furnished by certain islands of the East Indian Archipelago. In shallow waters off the coast of Timor corals are building extensive reefs. Similar reefs, strikingly fresh but entirely dead, extend from the littoral zone to the higher ground above the reach of the highest tides; and still others, showing various degrees of weathering, occur at different levels up to several hundred feet above sea level. A classic example of changes in land level is found in the temple of Jupiter Serapis built by the Romans near the seashore in the vicinity of Naples. The three columns left standing are bored by lithodomi to a height of 20 feet above the floor, and their shells remain in some of the holes. From this we infer that after the temple was built the land subsided more than 20 feet, carrying the temple into the sea, and that later there was uplift of about the same amount. This conclusion is confirmed by historic record.

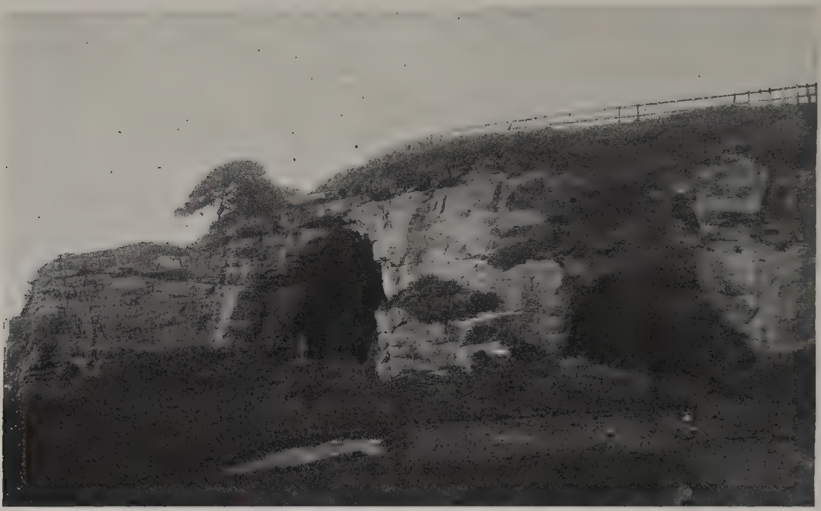


Fig. 184. — Ancient sea caves in former sea cliff at back of elevated beach, showing strand line. Coast of Fifeshire, Scotland. (Geol. Surv. of Scotland:)

Strong testimony is furnished also by the abnormal position of conspicuous features made by erosion and deposition along a coast. In parts of California, Chile, Scotland, and numerous other coastal regions raised beaches, accompanied by wave-cut and wave-built terraces, form nearly level benches of country terminated inland by former sea cliffs. Typical wave-formed caves pierce the cliffs (Fig. 184), and old stacks, now high and dry, rise abruptly from the terraces (Fig. 136). Such an elevated terrace, with its related features, is often spoken of as a raised *strand line*, since it commonly appears as a more or less distinct topographic line, or level, approximately parallel to the present shoreline and above it.

Still another kind of evidence, of a direct and positive character, is furnished by careful observations made year by year. Thus in countries bordering the Baltic Sea an uplift has been under observation for a long period, and has been measured by marks placed on the shores. In some places the elevation has been as much as 3 feet in a century, but the rate is not everywhere uniform. All the facts indicate that the Scandinavian peninsula has risen gradually for a long period, so that the northern part of Sweden is about 900 feet higher than it was at the close of the Ice Age. Raised strand lines are a noticeable feature in many northern regions. There is similar proof, on a large scale, that within recent geological time the west coast of South America has experienced very considerable elevation; and probably the uplift is still going on.

Depression. — Evidence of subsidence below sea level is less striking than that of elevation, but not less convincing. Submergence of features that are definitely characteristic of land surfaces constitutes the best proof. Submerged stumps and other marks of former forests are found at various places along the Atlantic coast from Maine southward. Some of the stumps remain rooted in the old forest soil, covered by marine muds or other deposits (Fig. 185). It is clear that this could occur only in situations, such as protected nooks and corners of estuaries, sheltered from the waves of the encroaching sea which would otherwise have swept away the forest soil. Submerged peat bogs, made under fresh-water conditions but now turned into tidal flats, are rather common and tell

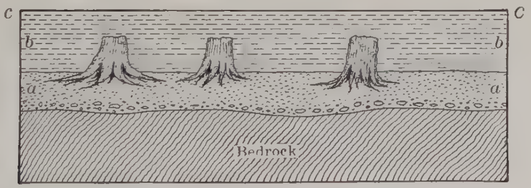


Fig. 185. — Showing submerged forest. *a*, old forest soil with stumps standing in it; *b*, marine deposits of silts and sands; *c–c*, present level of high tide.

a similar story. All the cumulative evidence goes to show that the Atlantic seacoast from Maine southward has gradually sunk within a late geologic period. Whether parts of it are still sinking is a matter about which geologists have not yet reached agreement.

Drowned Valleys. — Evidence of subsidence of the land is furnished by the irregular shorelines produced by the drowning of valleys, with production of bays and estuaries (Fig. 186). The seaward extension of river channels, such as the Hudson, for long distances across the submerged continental shelf, demands the same explanation; for manifestly these great trenches, now sunk in the sea floor, could not have been cut while the continental shelf was covered with water, but only by river or glacial action, or both, when it stood at a higher level and was a land surface. The submerged channel of the Hudson has been outlined by closely spaced soundings on the continental platform south of New York harbor.

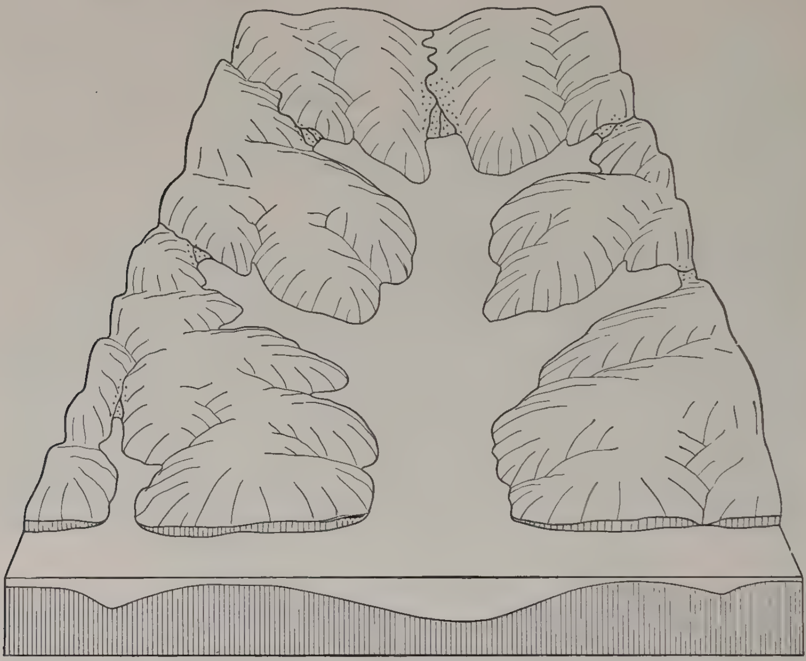


Fig. 186. — A river drowned in its lower reaches, forming a branching estuary. The estuary is slowly being destroyed by encroaching deltas.

Subsidence and Deposit of Sediment. — In many parts of the world thick deposits of sediment are being laid down by rivers in subsiding basins adjacent to coasts. Borings into deltas pass through alternating marine and fresh-water deposits, or even through sediments entirely nonmarine, to a great depth below present sea level. For example, wells sunk into the delta of the Po, near Venice, pass through four separate layers that contain abundant remains of plants similar to those now growing in the marshlands along the Adriatic. One of the layers is about 300 feet below sea level. At shallower depths some of the sands, gravels, and clays contain shells of marine molluscs, and other layers yield fresh-water types, such as land snails. In the delta of the Ganges, near Calcutta, pieces of wood and bones of land animals are found hundreds of feet below sea level. Deep wells in the Great Valley of California furnish similar evidence. Facts of this kind show that subsidence has been going on for a long period, not at an even rate, but as an interrupted process, whose variations permitted alternating fresh-water and marine deposits to be formed.

Evidences of Elevation or Depression Inland. — Movements of the lithosphere involving changes of level are not confined to the sea coasts; they occur also in the interior of the continents. For example, in 1811 large areas of the Mississippi flood plain near New Madrid, Missouri, sank far below their former level and now are occupied by lakes. Trees that grew on the plain were killed by the flooding, and for a long time their dead trunks or tops, projecting above the water, furnished evidence of the changes in level.

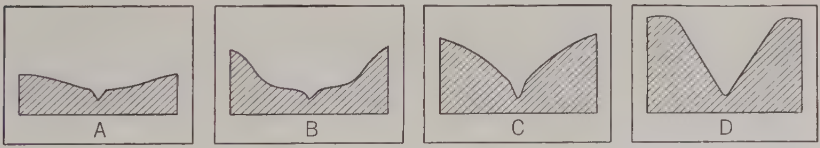


Fig. 187. — Cross profiles of valleys showing that the effect of uplift and rejuvenation depends on the stage at which it occurs.

A — Old stream rejuvenated: effect very strongly marked.
B — Mature stream rejuvenated: effect strongly marked.
C — Young stream rejuvenated: effect barely perceptible in the spurs along the valley sides.
D — Very young stream rejuvenated: effect not registered.

Rejuvenated Streams and Entrenched Meanders. —Upwarping of the crust is likely to increase the cutting power of the streams affected by increasing their gradients. Streams thus affected are said to be *rejuvenated.* The effects, however, upon the aspect of a valley will be far more noticeable in the case of an old or mature stream rejuvenated, than in the case of one rejuvenated when still in its youth (Fig. 187). This follows from the fact that renewed cutting normally trenches a new **V**-shaped valley in the floor of the old one, and that if the latter is already

V-shaped, its appearance in cross profile is scarcely altered. A keen eye can sometimes detect the evidence of uplift in a valley still youthful (Figs. 187 *C*, 188).

If a meandering stream is affected by uplift so rapid that the rate of lateral cutting by the stream cannot keep pace with it, the stream will be rejuvenated in the exact meandering course it followed immediately prior to the uplift. The meander loops thereby come to occupy gorges separated by steep-walled projecting spurs (Fig. 189). These loops are called *entrenched* meanders, and are a positive indication that uplift has taken place in the region where they occur. Furthermore, the depth of the gorges is an approximate gauge of the amount of uplift.

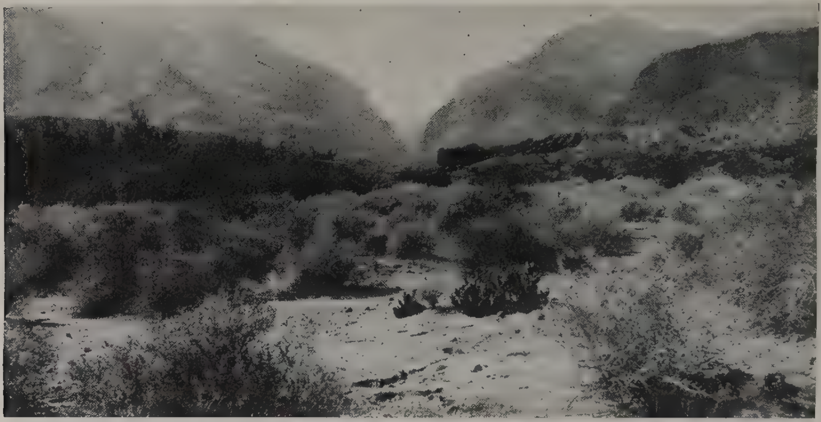


Fig. 188. — East end of Boulder Canyon, southern Nevada. Note abrupt changes in slope of valley walls, suggesting at least two rejuvenations. (U. S. Geol. Surv.)

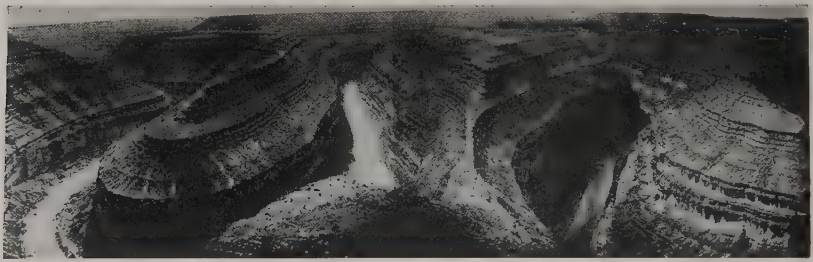


Fig. 189. — Entrenched meandering valley of the San Juan River, 30 miles below Bluff, Utah; canyon 1200 feet deep. (Vinson.)

Rock Terraces Caused by Successive Uplifts. — Rock terraces caused by differential erosion of horizontal strata are discussed in Chapter IV. Terraces are also not infrequently cut in tilted or completely folded strata, beveling them regardless of their unequal resistance (Fig. 190). Since hard and soft beds alike can be brought down to a common plane only near baselevel, it follows that each pair of rock terraces of this type must have been formed by a stream in the latter half of its cycle, and that for each pair of similar terraces there must have been a corresponding uplift of the land followed by renewed downcutting toward baselevel and then widening of the valley by lateral planation. Thus in Fig. 190, a nearly baseleveled surface *a* must have been uplifted and dissected, followed by valley widening near a new baselevel *b*. A second uplift resulted in the cutting of the plane *c* which after a third uplift was dissected into terraces by downcutting of the trench *d*. Obviously peneplanation took place in no case; if it had, the higher terraces would have been destroyed. Each terrace is veneered with a thin layer of alluvium, the remains of an alluvial flat built up by continuous deposition on the inside curves of meanders.

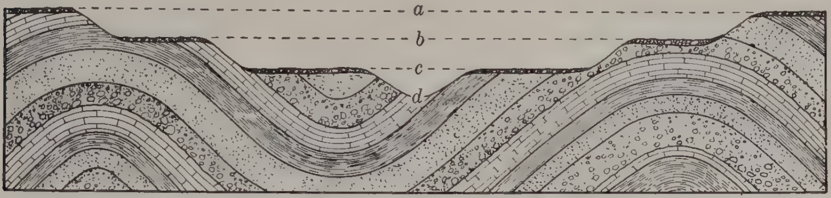


Fig. 190. — Rock terraces caused by successive uplifts. (After Wright. Va. Geol. Survey.) Terraces *a*, *b*, and *c*, are remnants of old graded flats, elevated by three successive uplifts, the last of which is causing the excavation of the trench *d*. Width of cross section is of the order of magnitude of $\frac{1}{2}$ mile. (Drawn by E. J. Lees.)

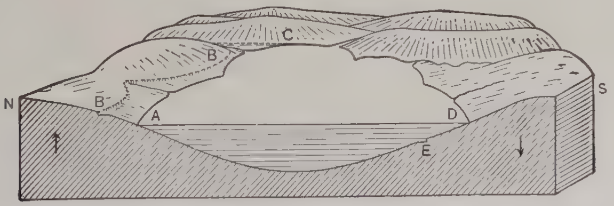


Fig. 191. — Showing tilting of lake basin. *AD*, present lake level. *BB*, raised shoreline disappearing under lake at *C*. Land has been raised on the north (*N*), and on the opposite side the former shoreline, *C* to *E*, has been drowned.

Tilting. — An excellent illustration of tilting on a large scale is afforded by the Great Lakes. To the northeast the land has risen since the retreat of the great ice sheet, and as a result the lake basins have been tilted southwestward. The old strand lines are several hundred feet above the present water level on the north and northeast, and slope toward it as they are followed west and south. Since the lakes discharge to the east, the raising of their outlets has caused them to enlarge, expanding them to the west and south. As a result, river mouths on the south and west sides of some of the lakes (especially Erie and Superior) have been drowned. The tilting movement is still in progress and has been accurately determined. It is at the rate of 6 inches per hundred miles per century. Small as this rate seems, in 1600 years it would cause the upper Great Lakes to discharge by way of Chicago River into the Mississippi drainage. (Fig. 191.)

Significance of Recent Changes in Level. — It is obvious that changes of surface level must be accompanied by bending or fracturing of the lithosphere. In fact, surface movements merely record deep-seated processes which lead to slight changes in the form of the globe. Commonly the movement is a broad warping that affects wide areas. Locally there is sharper bending, or the rocks actually break under the enormous strain, causing abrupt offsetting of roads or other features at the surface. Such breaks, or *faults*, are common along the Pacific coast of North America, in Japan, and in many other parts of the world. Displacement along some of these fractures within historic time has caused local shifts in level or horizontal changes of position.

RESULTS OF ANCIENT CRUSTAL MOVEMENTS

It is neither desirable nor possible to distinguish sharply between recent and older deformation. Certain features at the surface bear unmistakable witness to very recent crustal disturbance, although there may be no direct record of the event in human history. On the other hand we see in the rocks many fractures and folds that date from very early periods in Earth history, as is shown clearly by geologic relationships. Between these two extremes are found indications of disturbance in every geologic epoch, showing that movements of the same or similar kinds have been continuous or recurrent throughout recorded geologic time. In general the record of the latest movement is found in forms and features on the Earth's surface, such as the elevated beaches, drowned valleys, and tilted lake basins described above. All surface forms are ephemeral, because of erosion; and as they disappear or become fragmentary, the most reliable guide to former crustal movements is found in the structure of the underlying rocks.

Study of the framework or structure of the lithosphere is called *structural geology*. Such a study may be entirely geometrical, with the object of furnishing an exact description of the crust, including all breaks and bends in the rock units. Knowledge of this kind has great economic value, as it is essential in locating and following beds of coal or mineral veins. However, the structural features of the rocks have also a broader interest and value, as they furnish the clue to important events in the history of a region.

The principal structural features that indicate deformation of the lithosphere are broad bends or *warps*, *folds* of various sizes and degrees of intensity, and *fractures*.

Broad Warps

A permanent record of gentle warping is found in the bending or tilting of stratified rock formations that originally were nearly or quite horizontal. In southeastern Texas marine limestones and shales deposited in Cretaceous time now lie slightly above sea level, with a gentle inclination toward the Gulf. Formations of the same age, containing fossils of identical marine forms, are exposed in eastern Colorado, 5000 feet and more above sea level. When these beds were laid down a continuous seaway reached northwestward from the Gulf of Mexico across the United States. In later time this part of the continent was elevated by a broad warping movement, so that the formations in Colorado were lifted a mile higher than the corresponding beds in Texas, 700 miles away. In the Colorado Plateau of Utah and Arizona a thick blanket of marine strata lies thousands of feet above sea level, and is dissected by deep canyons. In a general way these strata are nearly horizontal; but if any one layer is followed and mapped in detail it is found to bow gently upward into irregular domes and bend downward into shallow basins. It is clear that this wide area of sedimentary beds, many hundreds of miles across, was not lifted up with absolute uniformity, but was bent somewhat, so that each layer now resembles a wide board that has been warped by exposure to the weather. The original surface of the uplifted mass has been entirely destroyed by erosion; but a record of the distortion is preserved in the form of each layer of stratified rock beneath that former surface.

The value of widespread marine sediments as a means of detecting such gentle warping of the crust is obvious. In areas occupied entirely by granites and similar massive rocks, a broad warp is recorded only in the surface forms; and after these forms have been obliterated by erosion there is nothing to suggest the ancient movement.

Folds

Anticline and Syncline. — In many places the stratified rocks have been buckled into plications or folds. Some of these are on a small scale and can be seen directly; but commonly the folding is on such a great scale, and exposures of the rocks are so discontinuous, that it is necessary to study and piece together the structure of certain distinguishable layers over many miles before the form of the folds becomes clear. The nature and scale of the folds in mountain regions can be appreciated best by study of maps and cross sections that have been prepared by geologists in the field.

Two terms are used constantly in descriptions and discussions of folds. Like regular swells on the ocean, rock folds ordinarily occur in a series, with alternating crests and troughs. The crests of the folds — that is, the upfolds — are *anticlines;* the troughs, or downfolds, are *synclines* (Fig. 192). Even if the original surface crests should be carried away by erosion, and the whole reduced to a level plain, we should still call the upfolded portions below the surface anticlines, the downfolded portions synclines, and in imagination reconstruct the missing parts (Fig. 193). Thus it should be clear that anticlines and synclines *are not a matter of surface topography, but of structure* (Figs. 194 and 195). Commonly the original configuration of the surface is even reversed by erosion, so that valleys now occupy the positions of the ancient crests, and ridges or mountains are in the places of the troughs; but the original structural terms still apply (Fig. 193).

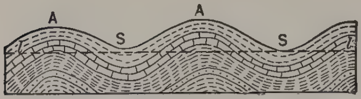


Fig. 192. — To illustrate anticlines, *A*, and synclines, *S*.

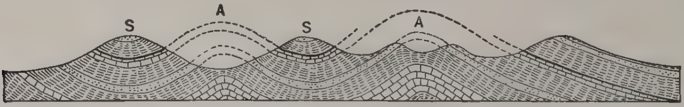


Fig. 193. — Anticlines, *A*, and synclines, *S*. Folds have no relation to present surface forms.

Outcrop. — Only the ideal relation of simple, upright, regular folds has been considered above. A series of folds approximating this form is by no means uncommon in nature; but usually the folding is much more complicated. Moreover some important facts about folds cannot be represented in ordinary cross sections. The varied kinds of deformation which the rocks have suffered in any region condition the *geologic structure* of that region; and it is a matter of the highest importance that the geologic structure of every country should be known so far as possible and represented accurately on maps. If the surface of the Earth were everywhere naked bedrock, this would be relatively an easy matter; but since the rocks have been greatly eroded, and are largely covered with earth and vegetation, or with water, snow, and ice,

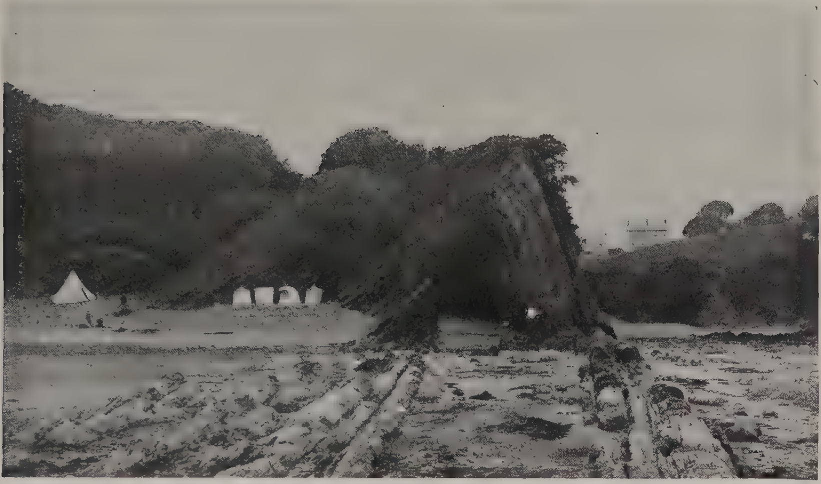


Fig. 194. — An anticline broken at the top. In the foreground the outcrops of the eroded strata are seen dipping outward, from which the anticline structure could be inferred if the arch did not exist. Pembroke, Wales. (Geol. Surv. of England and Wales.)

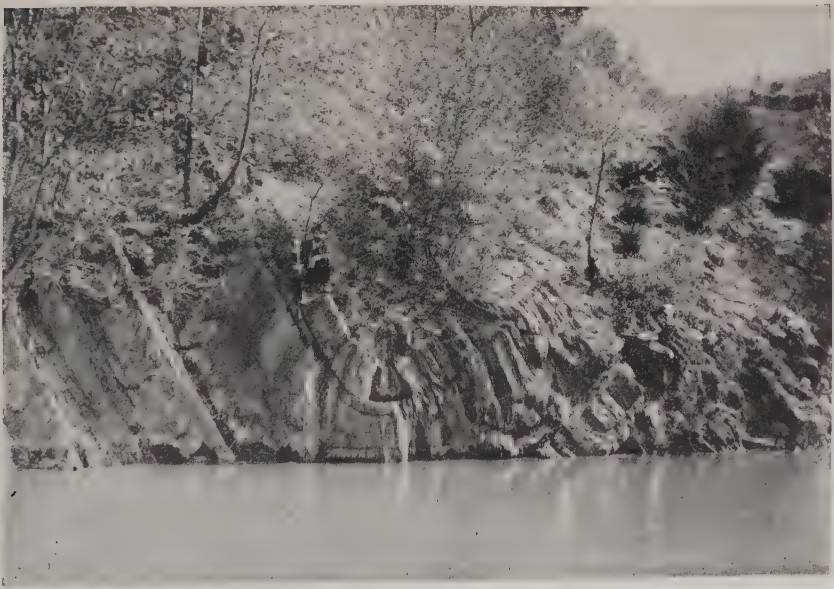


Fig. 195. — A syncline, near Hancock, Md. (U. S. Geol. Surv.)

the natural difficulties of the task are enormous. The structure in a region is determined by a careful study and comparison of the *outcrops,* by which term is meant the actual exposures of bedrock at the surface. If the ground were perfectly level and the strata horizontal, the outcrop would be the flat surface of the uppermost rock stratum, and we should learn little from it; but on slopes and cliffs bordering stream valleys, we may inspect the outcropping edges of many strata. If the sides of the valley were trenched by ravines, the line of outcrop would not be straight, but sinuous, retreating from the valley into the ravines, and advancing on the spurs (Fig. 196).

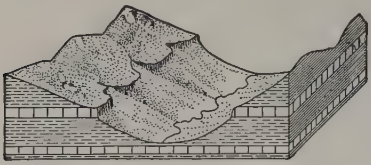


Fig. 196. — Section and outcrop of horizontal strata along valley.

If the strata have been inclined by folding and eroded their edges may be exposed even on a nearly flat land surface (see Fig. 194). Commonly the edges of the harder, more resistant beds project to form the more prominent outcrops. In mountain regions, soil and other concealing débris usually decrease in amount with increasing height; and exposures of rock grow in prominence correspondingly, until the upper rocky ridges and peaks may each be a vast outcrop. Because of the excellent exposures, and the great depth of the section visible in canyon walls and on the cliffy slopes, mountains furnish the most favorable opportunities for determining geologic structure.

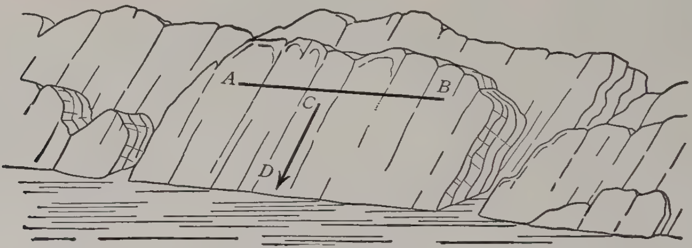


Fig. 197. — To illustrate strike and dip of tilted strata. The line of intersection of the horizontal water surface with a bedding plane gives the strike; or any other horizontal line on the bedding, as *AB*, also represents the strike. The line *CD*, which is at right angles to *AB*, indicates the dip of the layer.

Dip and Strike. — These terms, used constantly in describing the attitude of inclined strata, may be defined as follows: *Dip is the angle of inclination of the plane of bedding from a horizontal plane. Strike is the direction of the line of intersection of the plane of bedding with a horizontal plane.* Reference to a diagram will make the definitions clear.

In Fig. 197 any horizontal line, as *AB*, drawn on the surface of the inclined stratum represents the strike. *CD*, at right angles to the strike line, represents the maximum inclination of the surface, or the *dip*. The practical use of strike and dip are better illustrated in Fig. 198. A dark stratum, *AG*, is exposed on a perfectly flat surface. The direction of its outcrop, *PG*, is the strike. If this direction, measured by the angle *NSG* (= *NOK*), is 30 degrees east of north, the stratum is said to strike N. 30° E. Obviously a layer with this strike might dip either toward the northwest or toward the southeast. The stratum in the diagram inclines southeastward, making an angle *APH* with the horizontal. If this angle measures 40 degrees, the stratum is said to dip 40° southeastward.

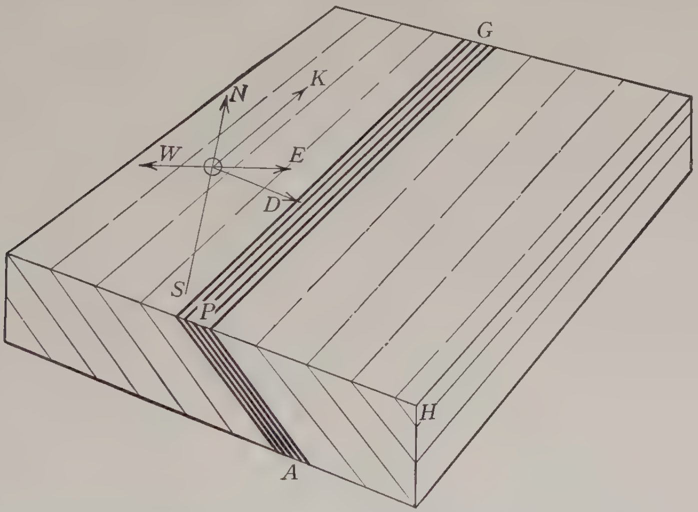


Fig. 198. — To explain strike of strata, and direction and amount of dip. The stratum, *A–G*, is exposed on a horizontal surface. The direction of the strike, with relation to the compass points NESW, is shown by the angle *NOK*, which is east of north. The stratum dips in the direction *OD*, which is east of south by the amount of the angle *SOD*. The angle *APH*, measured from the horizontal plane, is the angle of dip.

The direction of strike is taken with an ordinary compass, and the direction of dip is calculated. The *amount* of dip is taken with a clinometer, which is essentially a pendulum swinging over a graduated arc (Fig. 199). For geologic purposes the compass and clinometer are usually combined in one instrument.

Dip and strike are represented on geologic maps by a conventional sign ⊤, in which the direction of the cross bar, as placed on the map, indicates the direction of strike, and the arrow points in the direction

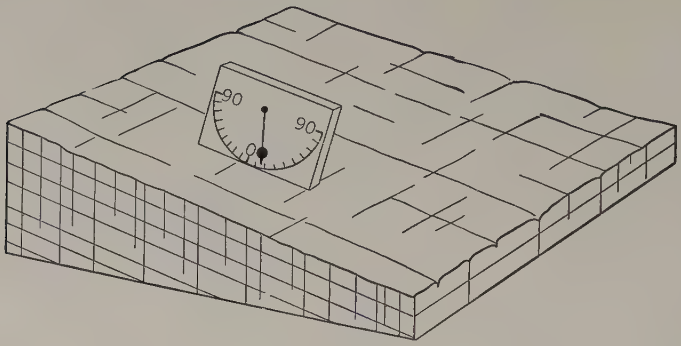


Fig. 199. — To explain measurement of dip angle with the clinometer. The pendulum of the instrument swings freely on an axis, and therefore is always vertical when the box is on edge. When the edge of the box rests on a bedding plane in the direction of dip, the angle of dip is read directly on the graduated arc.

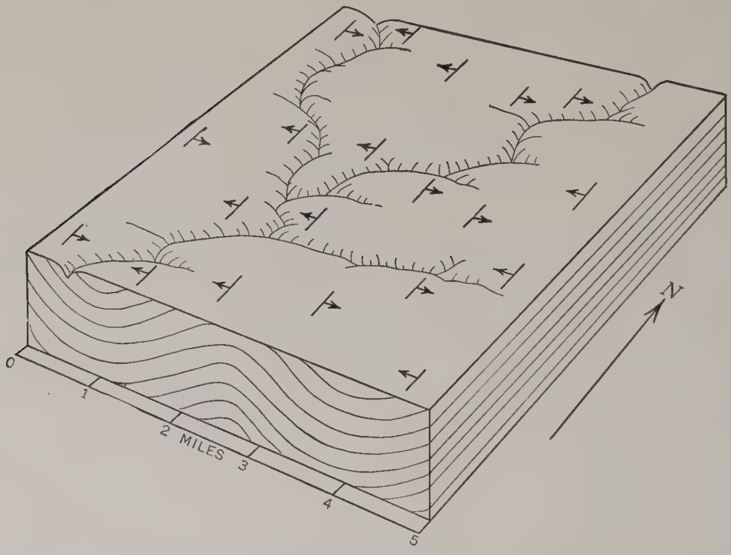


Fig. 200. — Diagram of a land surface underlain by folded strata. In making a geologic map of this area, the strike and dip of each outcrop is noted in the proper position, by use of the conventional symbol ⊤. The symbols in the figure indicate the positions of an anticline and two synclines, which are shown in section at the end of the block.

of dip (Fig. 200). Ordinarily the actual amount of dip in degrees is written in; thus, |→ 30°.

Pitching Folds. — A series of dips and strikes arranged on a map as shown in Fig. 200 indicates a set of parallel folds in the underlying rock. It is quite evident that folds of this kind could not run in the direction of strike indefinitely, or around the world; they must end somewhere. The end of a syncline, as seen on a map and in section, has the appearance shown in Fig. 201, *A*. A stratum is warped into a form like the end of a boat. An anticline, near its termination, resembles a boat overturned (Fig. 201, *B*). In either of these structural features the outcrop of a stratum, on a nearly flat surface of erosion, turns in the shape of a horseshoe; but there are two general ways for distinguishing one form of fold from the other. In the ordinary syncline, dips are consistently toward the inside of the horseshoe (Fig. 202, *A*), and the younger strata lie inside. In an ordinary anticline, dips are outward at every point, and the older strata lie inside the horseshoe (Fig. 202, *B*).

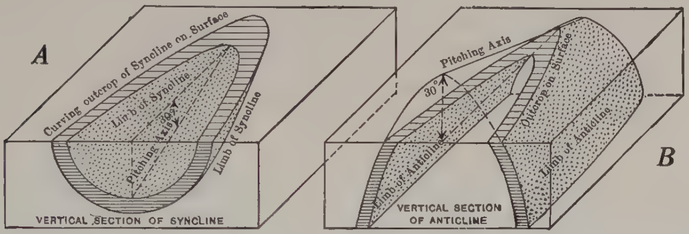

A
Curving outcrop of Syncline on Surface
Limb of Syncline
30°
Pitching Axis
Limb of Syncline
VERTICAL SECTION OF SYNCLINE
Pitching Axis
30°
Limb of Anticline
Outcrop on Surface
Limb of Anticline
VERTICAL SECTION OF ANTICLINE
B

Fig. 201. — *A*, ending of a syncline, as seen on a flat surface. *B*, near the ending of an anticline. For simplicity, only a single stratum is shown in each case. Consider that each block is about half a mile wide. (H. H. Robinson.)

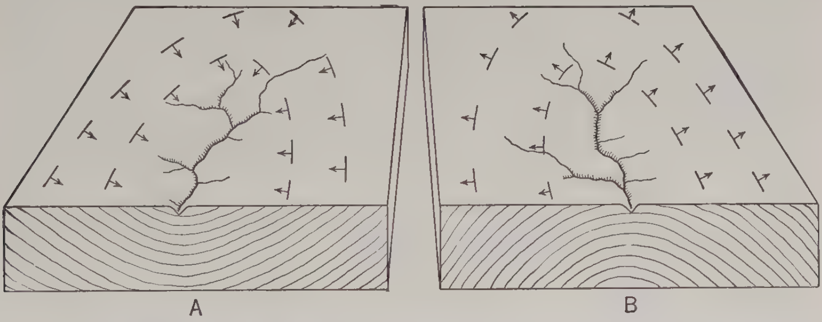


Fig. 202. — Illustrating the use of dip and strike symbols on a map to show a plunging syncline (*A*) and a plunging anticline (*B*).

The median line of a fold, along the top of an anticline or the bottom of a syncline, is the *axis*. This line extends along a bedding surface, or along this surface restored if it has been partly eroded (Fig. 201). The axis *emerges* from the ground toward the end of a syncline, *plunges* into the ground toward the end of an anticline. The angle between the axis and the horizontal is called the *pitch* of the fold. It is evident that the pitch is merely a special case of dip measured along the axis. In a *pitching* or *plunging* fold, the sides or *limbs*, as exposed on a nearly level surface, cannot be parallel, but necessarily converge or diverge (Figs. 201–203). If outcrops of strata in the two limbs run along parallel to each other for a long distance, the axis of the fold must be essentially horizontal. In the Appalachian region, hard sandstones or other resistant formations in the limbs of eroded folds make some of the high mountain ridges. Commonly two such ridges, on opposite sides of a great fold, maintain a straight parallel course for many miles; but eventually they converge and unite at the end of the fold.

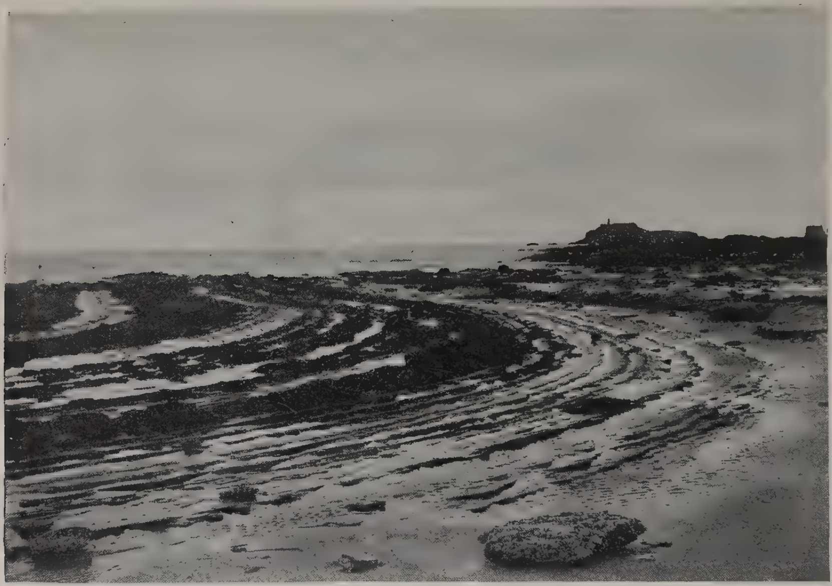


Fig. 203. — On a plain of marine erosion the outcropping edges of the strata are seen at the ending of a syncline, as shown by the curving strike and inward dip. Near North Berwick, Scotland. (Geol. Surv. of Scotland.)

Inclined, Recumbent, and Asymmetric Folds. — Thus far we have considered only simple, regular, upright folds. If a plane is imagined to pass through the center of a fold and its axis, as in Fig. 204, like the extended keel of a boat, we may call this the *axial plane* of the fold. In a regular or symmetric fold, this plane is one of symmetry; that is, the parts to left and right of it are symmetrically disposed, or each point on the left of the plane has its corresponding point at an equal distance on the right of it. If the fold is upright the plane is vertical (Fig. 204).

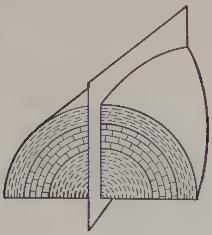


Fig. 204. — Upright symmetrical fold; axial plane vertical.

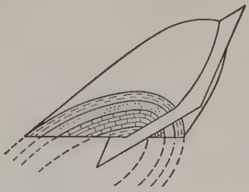


Fig. 205. — Inclined symmetrical fold; axial plane inclined.

However, some folds are not upright but have been pushed over until the axial planes are inclined. A fold of this kind is said to be *overturned* (Fig. 205). Such overturning may, indeed, go so far that the axial plane is nearly, or actually, horizontal; and the fold is then termed *recumbent*. Synclines may have a similar attitude.

It is also common to find that folds are asymmetric (without symmetry); that is, they are not similar to right and left of the axial plane, which is not, therefore, one of symmetry, as in a regular fold (Fig. 206).

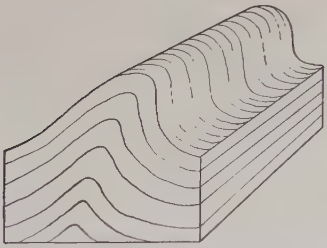


Fig. 206. — An asymmetric fold.

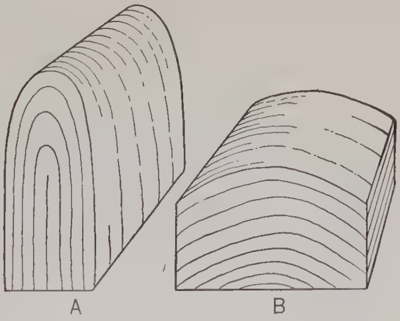


Fig. 207. — *A*, closed fold; *B*, part of an open fold.

Other Features of Folds. — If folds are so sharply flexed that the limbs are nearly or quite parallel, they are said to be *closed;* in this condition the horizontal distance across the strata, or the width of the fold, cannot be farther reduced without squeezing or mashing of the beds (A, Fig. 207). If the limbs make a large angle with each other (as in B, Fig. 207) the fold is *open* and the strata may be further folded without mashing.

Adjacent closed folds in which all of the limbs are parallel are called *isoclinal* (equally inclined) folds (Fig. 208).

Anticlines and synclines are the common types of folds and so have been given first consideration. A less common type is the *monocline*, in which the strata are bent in one direction only (Fig. 209). A true monocline is a one-limb flexure, on either side of which the strata are horizontal or have uniform gentle slopes.

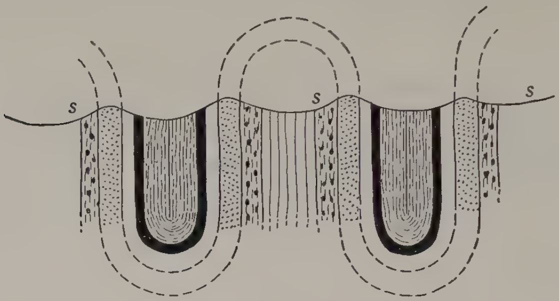


Fig. 208. — Isoclinal folds. *s–s–s* represents the present surface of the ground. Note that after the crests of such folds are eroded away the limbs form a series of parallel strata, and the fold structure is recognized at the surface only by the peculiar repetition of distinctive strata. Horizontal distance represented, two miles.

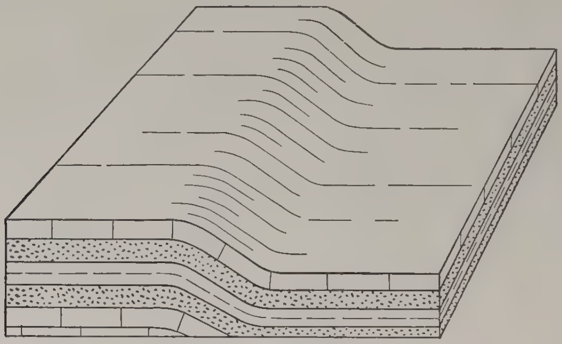


Fig. 209. — A monoclinal fold.

Geosynclines and Geanticlines. — Long belts within a continent or on the ocean floor have been warped down to form *geosynclines*, whose dimensions are measured in hundreds of miles. Correspondingly great upwarps are called *geanticlines*. The prefix in each case (from the Greek *Geos*, meaning Earth) emphasizes the scale of these features. In contrast with ordinary folds, the flexures responsible for geosynclines and geanticlines are very gentle. The Baltic Sea may be a modern geosyncline. In the region about Cincinnati, Ohio, the Paleozoic strata are bowed gently into a geanticlinal arch 250 miles wide.

The geosynclines of the past, as well as those of the present, have been the great basins for the accumulation of sediments, like those exposed in the Appalachians and the Alps. When later the accumulated beds are compressed into folds, the whole series may form a compound uplifted mass, which erosion carves into mountains.

Joints and Faults

In the outer shell of the Earth the rocks are traversed in all directions by fractures, varying from minute crevices to extensive breaks. We have considered the importance of fractures in the weathering of rocks and formation of soil, and in the circulation of ground water. They are, indeed, of great geologic importance, because of the processes that give rise to them and the results achieved by their aid.

A fracture on which there has been no appreciable displacement is called a *joint*. Ordinarily joints are closed so tightly that little or no space is visible between the walls. If the walls are distinctly separated, the term *fissure* is preferable to joint. Some fissures are open, and others have been filled with mineral matter deposited by circulating water. If there has been relative displacement of the walls in a direction parallel to the fracture, so that corresponding points on the two sides are distinctly offset, the fracture is known as a *fault*. Faults are very important geologic features.

Joints in Stratified Rocks. — Field examination shows that joints are common, but that they are much more numerous in some places than in others. Where they are abundant, commonly they are arranged in more or less definite *sets;* that is, the divisional planes running through the rock fall into groups according to direction. In many places there are two prominent sets of joints, approximately at right angles to each other and each set nearly vertical. Such a combination of two or more intersecting sets constitutes a joint *system*. Combined with natural divisional bedding planes, a well-defined system of joints divides stratified rocks into series of closely fitted blocks. The finer the grain of the rock, as a rule, the more perfect the jointing and the more definite the resulting blocks. Thus, in shale beds and in limestones, the jointing may be very perfect, as illustrated in Fig. 210.

Such jointing may result from various forces; for example, from the tension produced in the beds of sediments by contraction when they are elevated from the sea-bottom to form land masses, and undergo a drying-out process. A more common probable cause of regular fracturing is the warping and twisting suffered by the strata during crustal movements. In regions where the strata have been definitely folded, as in many mountain zones, unquestionably the compressive force that plicated the beds produced many joints. More locally, and especially at the crests of anticlines, joints may have been made by stretching or tension.

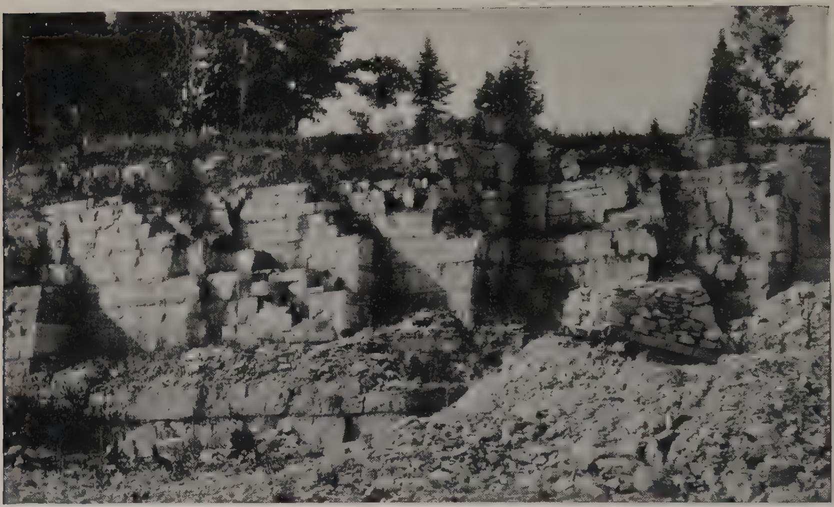


Fig. 210. — Illustrating joints in limestone beds. The horizontal lines are bedding planes. There are two sets of joints, nearly vertical and at right angles to each other. Drummond Island, Mich. (U. S. Geol. Surv.)

Joints in Igneous Rocks. — A common type of jointing in igneous rocks is due to the contraction resulting from the cooling of the original magma. This occurs during and just after solidification from the liquid state. The character of the jointing depends on the rate of cooling, the size and shape of the igneous body, and other factors. Thus intrusive masses of granite and similar rocks are characteristically cut by joint planes that divide them into large blocks or prisms. Some of the smaller masses, such as sheets, laccoliths, and dikes are divided into small angular fragments by closely-spaced joints. In some laccoliths and similar dome-shaped intrusions there is a shelly jointing on a large scale, parallel to the domed surface. This appears to have been caused by nearly uniform cooling of the mass from the periphery, with resulting separation into sheets.

The most striking kind of contraction jointing in an igneous rock results in development of *columnar structure.* This result is produced, in general, when two dimensions of the mass are great and the third is small, as in a dike, a sill, or a lava flow. Closely fitted prisms, which are subdivided by inconspicuous cross joints, are developed in such a mass (Fig. 211). The prisms have a variable number of sides, but commonly they tend to be hexagonal, and some of them have remarkable regularity of form. They range from several inches to a number of feet in diameter, and up to 200 feet or even more in length. The Giant's Causeway on the north coast of Ireland is one of the most celebrated examples of this columnar structure. The columns form at right angles to the chief cooling surfaces, and consequently in a flat intruded sheet or flow of lava they stand vertically, whereas in a vertical dike they tend to be horizontal. Thus some dikes, exposed as walls by erosion, resemble regularly piled cordwood. Columns in volcanic necks tend to have a radial arrangement, like a great fan.

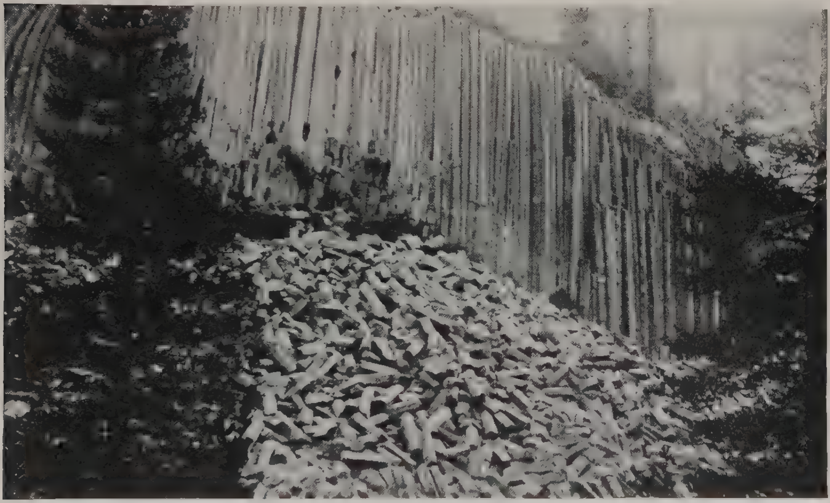


Fig. 211. — "Devil's Post-pile"; columnar jointing in lava. Head of San Joaquin River, Calif. (U. S. Geol. Surv.)

The reason for columnar structure appears to be that in a cooling mass, centers of contraction tend to occur on the cooling surface at equally spaced intervals. From each center three cracks form and radiate outward at angles of 120°. Intersection of these cracks produces a regular hexagonal pattern, and their penetration inward makes the columns. But nature is complex, and the ideal pattern is commonly modified by the occurrence of five- and four-sided figures of varying dimensions. By contracting lengthwise the individual columns break into sections. The same principles of contraction result in the polygonal shapes commonly seen on mud flats that have cracked from drying (Fig. 146).

Practical Importance of Joints. — Joints are a matter of great importance in all quarrying, tunneling, and mining operations where rock work enters as an important factor, since the jointing obviously facilitates progress. Without them, every rock fragment would have to be broken or blasted loose from bedrock. However, joints may also be a serious inconvenience, especially if large blocks of quarried stone are desired. Perfect monoliths 50 or 100 feet in length can be obtained from comparatively few localities.

General Features of Faults. — Displacement of rock masses along a fracture occurs either at the time of the original break, or at some later time. Faults are common features in rocks of all kinds. They are most evident in stratified formations, as the offsetting of layers makes the break conspicuous and directly measurable. However, faults affect massive igneous rocks as well; and since mineral veins or other features of economic value are displaced by such fractures, it is important from a practical as well as a scientific standpoint that the nature of faults be well understood.

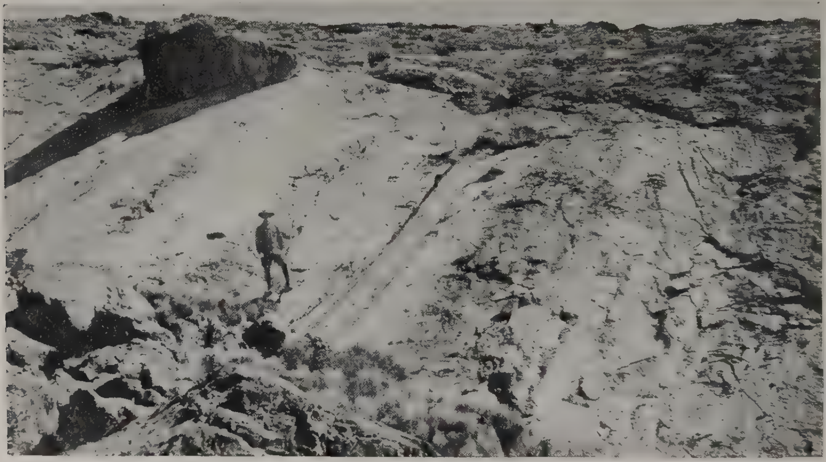


Fig. 212. — Part of an old fault surface, with slickensides, uncovered by erosion long after the original scarp was eroded away. Note the lined and fluted character, indicating that the movement was directly down the dip of the surface. Spotted Range, Nevada. (Longwell.)

The surface of fracture along which movement and dislocation has occurred is spoken of as the *fault surface*. However, the movement in faulting very commonly occurs upon a number of more or less closely adjacent breaks, producing a *fault-zone,* in which the various offsets make in the aggregate the total displacement. Such a distribution is sometimes called *step faulting.* The masses of rock involved in fault movements are generally of such size and weight, and so compressed together, that the motion of one fault face on the other takes place under tremendous pressure. As a result of the friction, the rock faces are smoothed and striated, and receive a high polish. Such polished and grooved surfaces are known as *slickensides* (Fig. 212). The line of intersection of the fault with the plane of the horizon is called the *strike,* or trend, of the fault, just as we speak of the strike of upturned strata. The surface of faulting is rarely exactly vertical; commonly it is inclined, and in some important faults it approaches horizontality. The angle between the fault surface and the horizontal plane is the *dip.* In an inclined fault the side that overhangs is known as the *hanging wall,* the other as the *foot wall* (Fig. 213). If one were to descend along a fault, as in an inclined shaft of a mine, the appropriateness of these old mining terms would be evident.

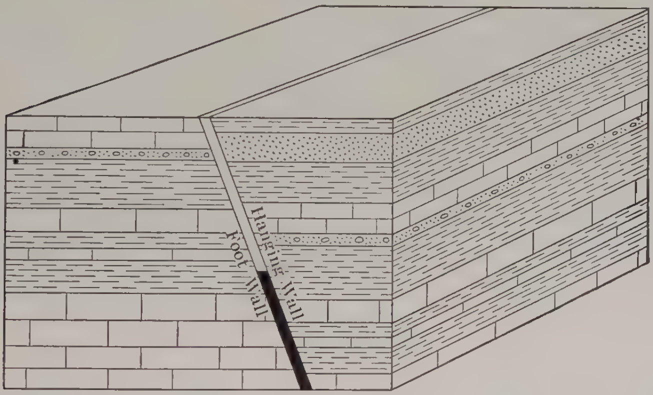


Fig. 213. — To explain fault relations and terms. The strata have been displaced by a fault, and a vein of mineral (black) has formed along the fracture. Mining operations have removed the mineral to a considerable depth, exposing the hanging wall and foot wall of the fault.

Generally the fracture is closed tightly; but parts of it may have been open at one time, and have been filled with mineral matter deposited from solution. Along many faults the grinding of the walls upon one another has produced a zone of broken and crushed rock known as *fault breccia.* Commonly there is a thin seam of clay-like material, known as *gouge,* directly along the fault. In the displacement of stratified rocks the friction usually causes bending of the layers near the fault surface. This feature, referred to as *drag,* may be a useful aid in determining the relative direction of motion on the two sides of the fault (Fig. 214).

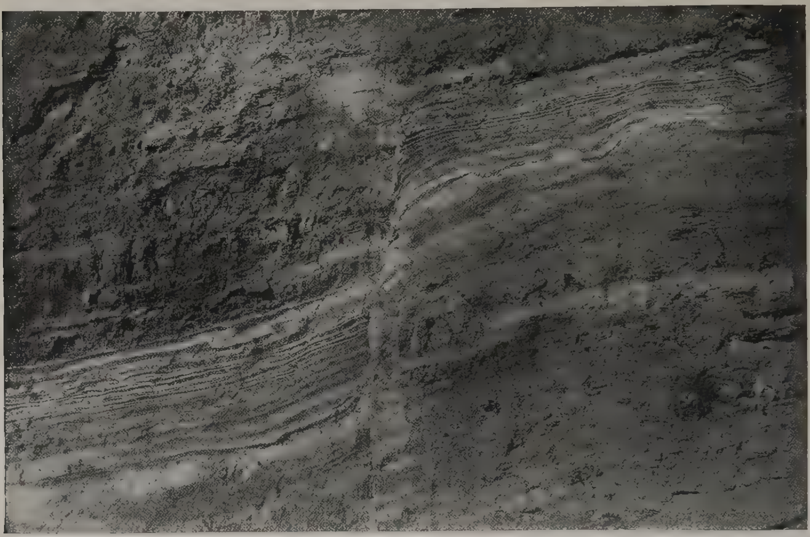


Fig. 214. — Fault in shale; the drag of the beds shows that the left side has gone down, the right up. Little River Gap, Tenn. (U. S. Geol. Surv.)

The features explained so far have to do chiefly with faults as seen below the surface of the ground. Ordinarily a fault breaks the surface as well as the rocks beneath; and if one side of the fault is elevated with relation to the other, the result is an abrupt cliff, or *fault scarp.* With the passage of time the original scarp is modified or even removed by erosion.

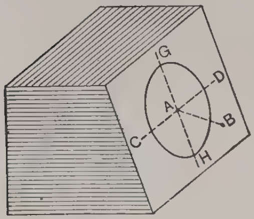


Fig. 215. — To show possible motion in faulting. The hanging-wall block is omitted, to show entire fault surface.

Motion on the Fault. — If we assume that one side of a fault stands fast, motion on the other side is either vertical, horizontal, or oblique. Thus in Fig. 215 the lettered plane represents one fault face, which for convenience is considered to remain at rest. It is exposed by removal of the block which is assumed to have moved. If we suppose that some particle at *A*, for example a crystal, was cleft by the fault, then one part *A* remained in its original place and the other part, embedded in the opposite face, was carried in some direction by the faulting. Suppose that the moving portion of the crystal scratched a groove on the stationary fault face. Depending on the direction of movement, this scratch or *striation* might be one of the lines *AH*, *AG*, *AC*, or *AD*. Most commonly it would be in some oblique direction, as *AB*.

Normal and Reverse Faults. — If faulting takes place by movement upward or downward, two different kinds of structure may result. In *A*, Fig. 216, the hanging wall has apparently slipped down with reference to the foot wall; a fault of this kind is known as a *normal* fault. In *B* of this figure, the hanging wall has apparently been crowded up over the foot wall; a fault of this kind is called a *reverse* fault. In the normal movement a particular layer *V–V* has been lengthened apparently by an amount corresponding to the gap *C–E*; in the reverse movement it has been shortened by an equivalent overlap.

It should be realized that the movement of one wall or the other to produce a normal or a reverse fault is purely relative. Thus a normal fault may result from keeping the hanging wall stationary and moving the foot-wall block upward; or both blocks may move, in opposite directions. In nature we see only the effect of faulting, and the actual character of the movement must be inferred.

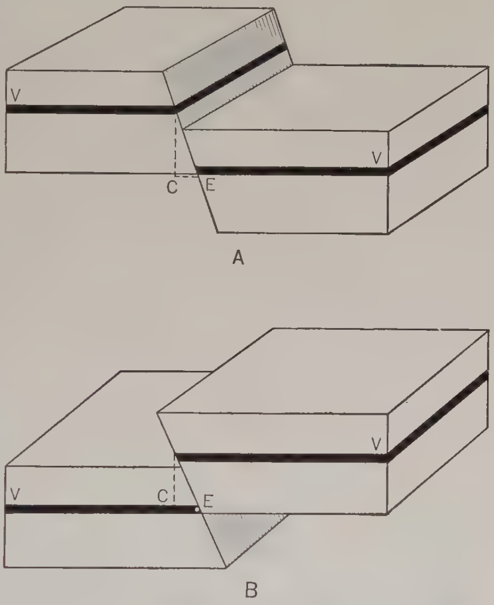


Fig. 216. — *A*. Simple normal fault. *B*. Simple reverse fault.

Components of Faulting. — It is often necessary for purposes of description or measurement to resolve a fault into component parts. On the diagram, Fig. 217, *AGEF* represents the horizontal plane and *A–F* is the strike of the fault. Let us suppose that the motion has been such that a particle near *A* has been carried to the position *B*; then the line *AB* joining these two positions represents the *displacement*, or *slip*; no matter what actual path the particle may have followed, *AB* is the resultant, and its length the measure of the slip. The line *AB*, however, in order that it may be fixed and determined, must be referred to known axes or planes. The line AC gives the amount of motion in the horizontal direction at right angles to the strike FA, and this is known as the *heave* of the fault; the line CH is the amount of vertical motion and is called the *throw* of the fault; HB ($= AM$ or CD), the amount of motion in the horizontal direction along the strike, is termed the *strike-slip* of the fault. Thus there are the three right-angled axes AC, DC, and HC, meeting in the common point C, and these may be termed the *component axes* of faulting. The directions and intercepts on these axes being known, the displacement can be calculated, and the problem of the fault solved.

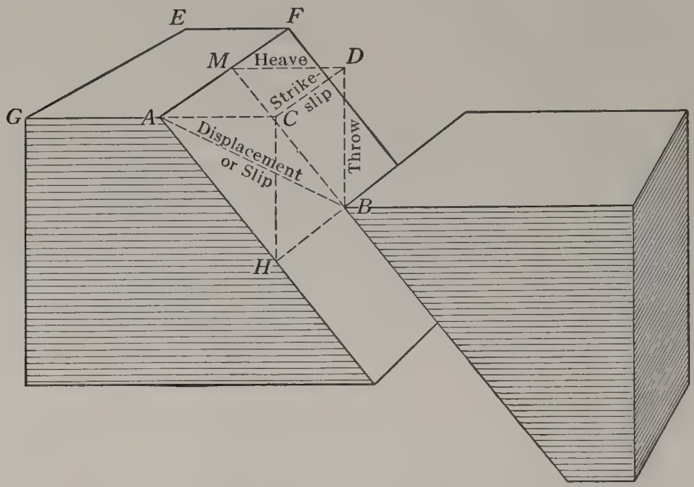


Fig. 217. — To illustrate and define the components of a fault.

Faults in Stratified Rocks. — Although faults occur in all kinds and combinations of rocks, they show to best advantage in stratified beds, on account of the strongly marked stratification which they disarrange. Certain terms are used to define faults in relation to the structure of the sedimentary beds. Thus in a *strike fault,* the strike of the fault and that of the strata are parallel, or nearly so, as illustrated in Fig. 218; *dip faults* cut directly across the strike of the strata, or nearly so, as shown in Fig. 219; *oblique faults* cut diagonally across the strike of the strata. The figures show only normal faults; but reverse movement produces similar effects. The figures also indicate no real strike-slip, but in Fig. 219, C there is offsetting of beds, with a false suggestion of strike movement. Such abrupt offsetting of tilted strata is one of the surest indications of dip or oblique faulting. Strike faults are more difficult to perceive and may easily be overlooked; they may cause

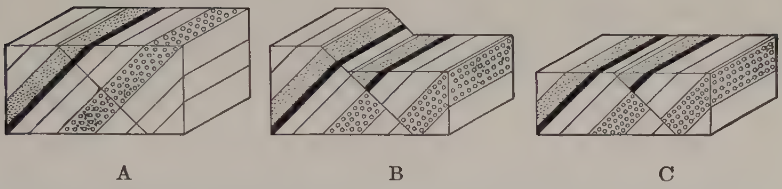


Fig. 218. — Illustrating strike faulting in stratified rocks. *A*, before faulting; *B*, after faulting, fault scarp still uneroded; *C*, surface levelled by erosion.

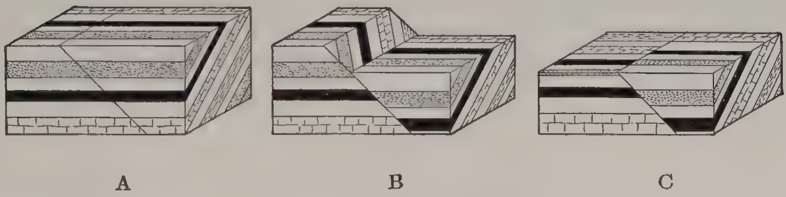


Fig. 219. — Illustrating dip faulting: *A*, before faulting; *B*, after faulting, fault scarp uneroded; *C*, surface levelled by erosion, showing offsets of strata.

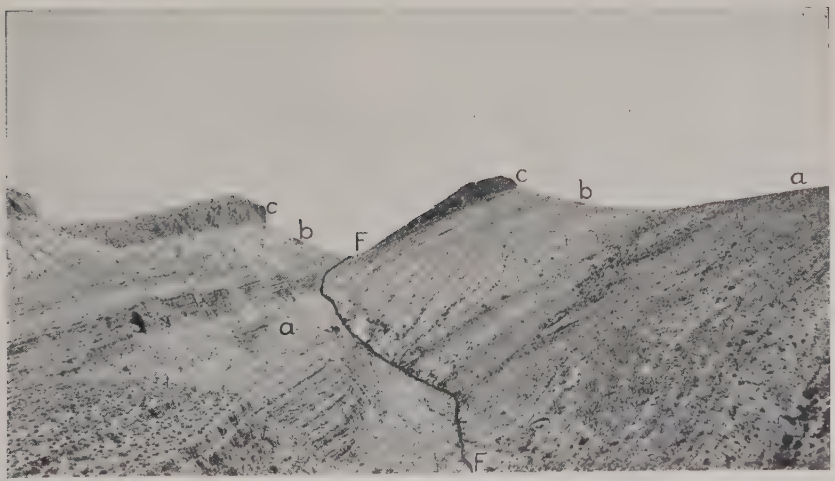


Fig. 220. — Repetition of formations by a normal strike fault (*F–F*). *a*, *b*, and *c*, are three distinct formations, broken and offset as shown. Erosion has removed much of the rock on both sides of the fault since the movement occurred. Spotted Range, Nevada. (Longwell.)

deception as to the thickness of strata by producing repetitions (Fig. 218, *C* and Fig. 220). Thus, in traversing strata the repetition of a certain set should lead to suspicion of strike faulting.

The movement of one side of a fault on the other side may be attended by a rotary or pivotal motion, as illustrated in Fig. 221. A fault of this nature is known as a *rotary* fault; or, if the displacement dies out gradually up to a definite point, it is a *hinge* fault. After erosion has levelled the surface, a fault of this kind is indicated by a pronounced difference in the strike and dip of strata on opposite sides of the break. Some faults pass gradually into monoclinal folds (Fig. 222).

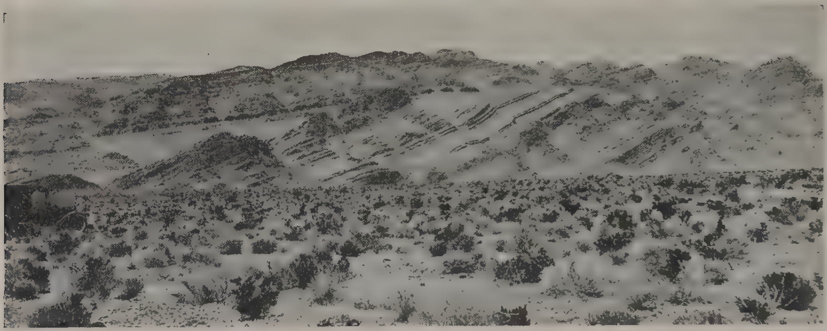


Fig. 221. — A pivotal or hinge fault. The fault runs from left to right, between tilted rocks and steep cliff on nearly horizontal strata. Displacement decreases toward the right, increases toward the left. Maximum throw about 3000 feet. Spotted Range, Nevada. (Longwell.)

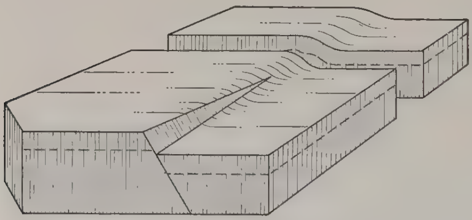


Fig. 222. — A normal fault passing into a monoclinal fold.

The Magnitude of Faulting. — The scale on which faulting has taken place varies within the widest bounds, from a fraction of an inch up to several miles. In the Plateau region of Arizona and Utah, several faults of great magnitude extend in a north–south direction, some of them crossing the Grand Canyon. Each of the largest fractures in this group can be followed 100 miles or more, and has a throw measured in thousands of feet. The Great Basin region presents the phenomenon of faulting on a colossal scale. In the area between the Sierra Nevada on the west and the Wasatch on the east, the crust is divided into huge blocks by gigantic fractures; and differential displacement of these blocks, together with erosion, has resulted in mountainous topography. A sunken tract of country due to downfaulting, or to uplift of adjacent areas, is called a *graben* (German, meaning ditch). Illustrations are the Jordan Valley and the Dead Sea, the great Rift Valleys of Africa, and Death Valley in eastern California. An upstanding mass between two grabens is a *horst.*

Thrust Faults. — Reverse faults are most common in those regions where crushing and folding of the Earth's crust have taken place; and the stronger the folding or crushing has been, the greater and more evident the reverse faults are. They are especially evident in the stratified rocks of mountain regions, as in the southern Appalachians and in the Alps. Careful and detailed study of old eroded mountain areas has disclosed reverse faults of tremendous displacement, some of them with a comparatively low angle of dip, or even quite horizontal. A reverse fault that has a gently inclined fault surface is known as a *thrust fault,* or simply a *thrust.* Many of these are of such magnitude and importance that they are commonly considered by themselves as a special class of faults. The surface on which movement occurs is spoken of as the thrust surface, or less accurately as the thrust plane.

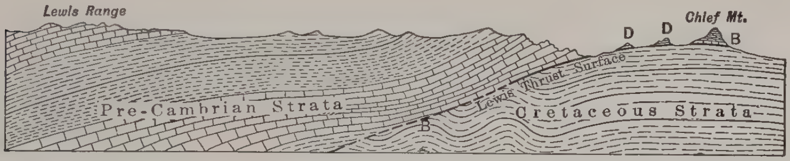


Fig. 223. — Section showing the thrust in northern Montana, whereby very old geologic formations of the pre-Cambrian are made to override the much younger beds of the Cretaceous. *BB* is the surface of thrusting; *D*, *D*, and Chief Mountain are erosional remnants of the pre-Cambrian resting on, and surrounded by, the younger Cretaceous. Displacement by thrusting observed, 7 miles; total amount unknown. (Generalized after Willis.)

Such thrusts have been discovered and studied especially in the Alps, in northwestern Scotland, in the Scandinavian peninsula, in the southern Appalachians, in the Rocky Mountains from British Columbia to Utah, in southern Nevada, California, and in many other regions. The horizontal displacement of lower, older formations over younger rocks ranges from several miles up to 25, 30 or even 40 miles for individual thrusts. Figure 223 represents a portion of the great thrust along the front ranges of the Rocky Mountains in northern Montana. The deciphering of these great displacements is one of the triumphs of modern geological research.

Topographic Results of Faulting. — If a fault of considerable magnitude were to be formed suddenly, it would naturally be marked by displacement of the Earth's surface, giving rise to a cliff, or *scarp.* Numerous fault scarps are recognized, and some of them have been formed within historic time.

Such scarps in their original form may be called *initial fault scarps.* As the process of weathering and erosion works more actively in general on the uplifted side, the scarps become dissected and lowered, and slowly retreat from the fault line. Thus they pass through youthful, mature, and old stages. Finally, the difference in elevation on opposite sides of the fault line may disappear completely, and thus all topographic expression, initially due to faulting, may be obliterated. This would finish one cycle of erosion on a faulted surface (Fig. 224).

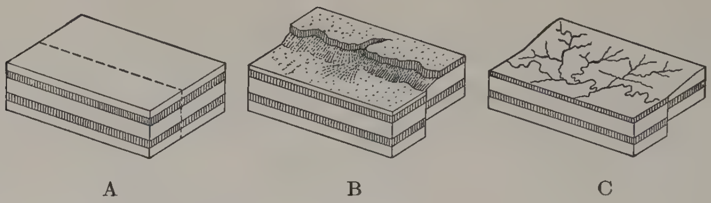


Fig. 224. — To show the origin, development, and possible history of an initial fault scarp. *A*, block of strata containing two harder, more resistant intruded sheets of trap, before displacement. *B*, after faulting and some erosion; the fault scarp has become mature, and has retreated from the fault line. *C*, approaching the end of the first cycle of erosion; the fault scarp has been obliterated.

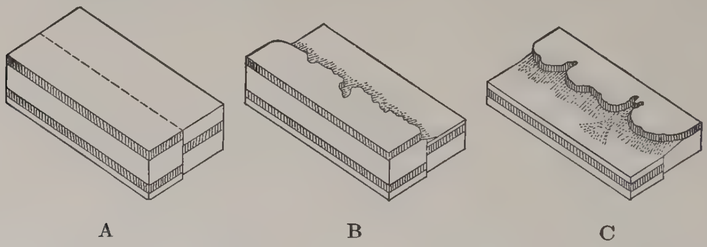


Fig. 225. — Possible development of fault-line scarps. *A*, the faulted block of the preceding figure commencing a second cycle of erosion; intruded trap sheets more resistant than the enclosing beds; uplifted block to the right. *B*, after erosion; a *fault-line* scarp has formed which faces *toward* the uplifted block. *C*, continued erosion has carried away the top trap of *B* and a new cliff has formed facing the other way, toward the sunken block; this fault-line scarp resembles the initial fault scarp (compare *B*, Fig. 224).

If now the whole region is uplifted, without further displacement on the fault, and a new cycle of erosion thus initiated, then the agents of erosion find on opposite sides of the fault line rocks of quite different hardness and ability to withstand their attack. Therefore one side may be lowered so much more rapidly than the other as to leave the latter standing in a cliff or escarpment (Fig. 225). As such a cliff results, not directly from the initial faulting movement, but from subsequent differential erosion, it deserves a distinguishing name, and has been termed by W. M. Davis a *fault-line scarp.* (See Fig. 225, which continues the history of the fault shown in Fig. 224.) The varying resistance to erosion on the opposite sides of the fault line determines naturally on which side the scarp will form, and some very long and old fault lines are marked in different parts of their courses by cliffs facing in opposite directions. Finally, through the completion of another cycle of erosion these cliffs in turn are worn away. The influence of an old fault that reaches deep into the crust is evident in the topography so long as erosion of the area is active. Determination of the exact history of any fault is possible only after very careful geological study.

The east slope of the Sierra Nevada and the west slope of the Wasatch Mountains are fault scarps that have undergone a large amount of erosion. Nevertheless the faulting is recent in a geological sense. At the base of each of these ranges some of the movement has occurred so recently that initial scarps may be seen almost uneroded in the soft fans of alluvial material brought down by the streams. The last recorded movement near the Sierra Nevada took place in 1872. These faults have grown by successive movements, and the upper part of each scarp has suffered most from erosion.

The Plateau region, through which Colorado River cuts its way, is crossed by a series of great faults which are marked by prominent cliffs. These have been described as actual fault scarps; but most of them are fault-line scarps developed during a second cycle of erosion. Examples on a smaller scale are very common. Thus the sunken tract of sandstone and intercalated trap sheets between New Haven, Connecticut, and Springfield, Massachusetts, is divided into a series of tilted blocks by faulting. It has passed through at least one cycle of erosion, in which the initial fault scarps were eroded away; it is now in another cycle, initiated by broad uplift of the region without further movement on the faults.

Origin of Faults. — The immediate cause of faults is comparatively simple and generally agreed upon; they result from strains set up in the outer shell of the Earth. Relief occurs by movement of rock masses, either along the surfaces of some previous fracture, or by the formation of a new one. Compressive strains give rise to reverse faults and thrusts; and it is natural that features of this kind occur commonly in areas of folded rocks, as folds also represent failure under compression. In regions of broad warping the rocks may be broken by torsion or twisting, which sets up tensional strains. After fracturing, displacement occurs by gravitative settling and readjustment of the fault blocks.

The *ultimate cause* of faulting and folding evidently depends on those processes within the Earth which set up strains in the lithosphere. These forces are most strikingly displayed in the formation of its chief features of relief, such as mountains and plateaus. The forces themselves are hidden and can be inferred only from their effects. As the subject is obscure at best, and speculation must be guided by consideration of all available facts, it is best to postpone inquiry into the ultimate cause of crustal deformation until the structure and history of mountains have been discussed.

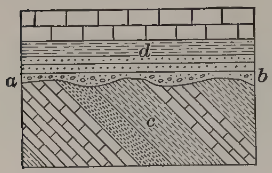


Fig. 226. — Section to show unconformity. The figure represents a vertical outcrop such as might be seen in the wall of a canyon. A conformable set of strata *d* rest unconformably upon another conformable set *c*; the line *ab* represents the unconformity.

UNCONFORMITY AND ITS MEANING

Definition. — In the diagram, Fig. 226, the layers of strata *d* have been deposited at one period and under one set of conditions; they are, therefore, spoken of as a *conformable* series of beds. Also beds of the series *c* are conformable among themselves; but it is quite evident that they are not conformable with *d*. Their attitude, and the abrupt termination of each bed upward, indicate that this series was tilted strongly and subjected to erosion before deposition of the overlying strata. The two series are *unconformable* with respect to one another, and the surface *ab* separating them is called an *unconformity*. It should be understood clearly that the unconformable contact is a widespread *surface* and not merely the *line* exposed in a vertical section. If the surface *ab* can be identified positively as a record of erosion preceding the deposition of overlying strata, it is a surface of unconformity. If the contact between rock masses of unlike character is due to faulting or to igneous intrusion, the contact surface is not an unconformity.

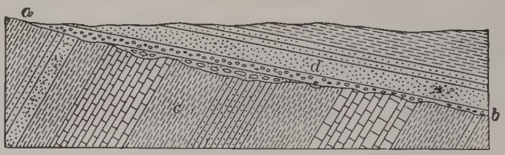


Fig. 227. — An angular unconformity with both series of beds tilted.

Geologic History Revealed. — Suppose that *c* and *d*, Fig. 227, are two unconformable series of marine strata. Then the relationships represented in the diagram indicate a definite succession of geological events. First there was a long period of quiet deposition in which the beds of set *c* were laid down in a horizontal position, or nearly so, on a sea floor. The thickness of the beds, the kinds of rocks (limestones and shales), and the contained fossils constitute the record for this period. At some time later than the deposition of the youngest beds in this series, there was strong folding or faulting by which the strata were tilted steeply. Possibly the deformed mass reached to mountain heights. At any rate the deformation was succeeded by erosion, which planed the upended strata to a nearly even surface. Therefore there is a gap or "lost interval" in the geological record at this locality. Next in the geological history there followed a period of subsidence, when the eroded surface became sea bottom again, and received a new deposit of sediments, forming the conformable series of strata *d*. The events of this time are recorded continuously in the strata as before. Finally, after a second period of uplift, tilting, and erosion, the whole record is presented to us to be read so far as the evidence permits.

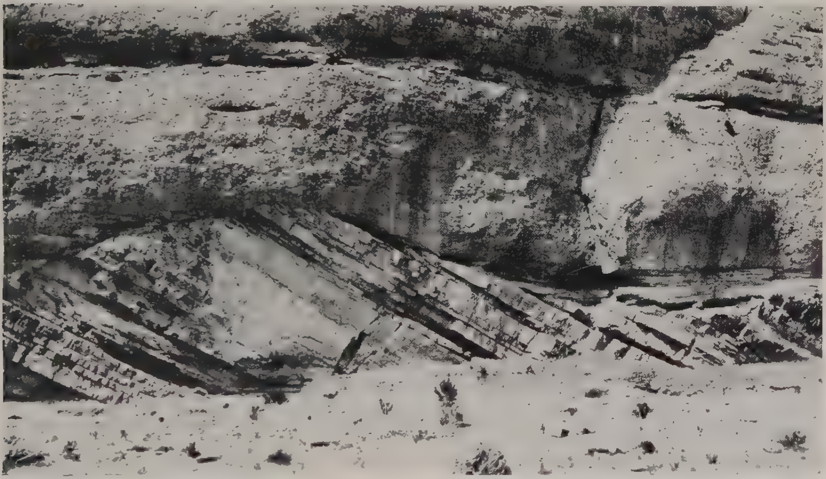


Fig. 228. — Angular unconformity between tilted and eroded Paleozoic limestones and horizontal Tertiary conglomerate. Meadow Valley, Nevada, along the Union Pacific R. R. (Longwell.)

Classification of Unconformities. — Unconformities may be divided into two main groups. In the first, the lower formation, either by the tilting of the beds or by its composition of non-stratified rocks, shows at once its nonconformity with the series of beds above it. It is called an angular unconformity (Figs. 228, 229, *C* and *D*) if the bedding planes of two stratified series meet at an angle. This term is good so far as it goes, but it does not cover the whole case, since the lower formation is not always composed of stratified rocks but may be of massive igneous or metamorphic rocks (Fig. 229, *A* and *B*). A more general term is needed, and an unconformity of this class is here termed a *nonconformity.*

On the other hand, the lower formation may be elevated, eroded, and submerged without material disturbance of the position of the beds.

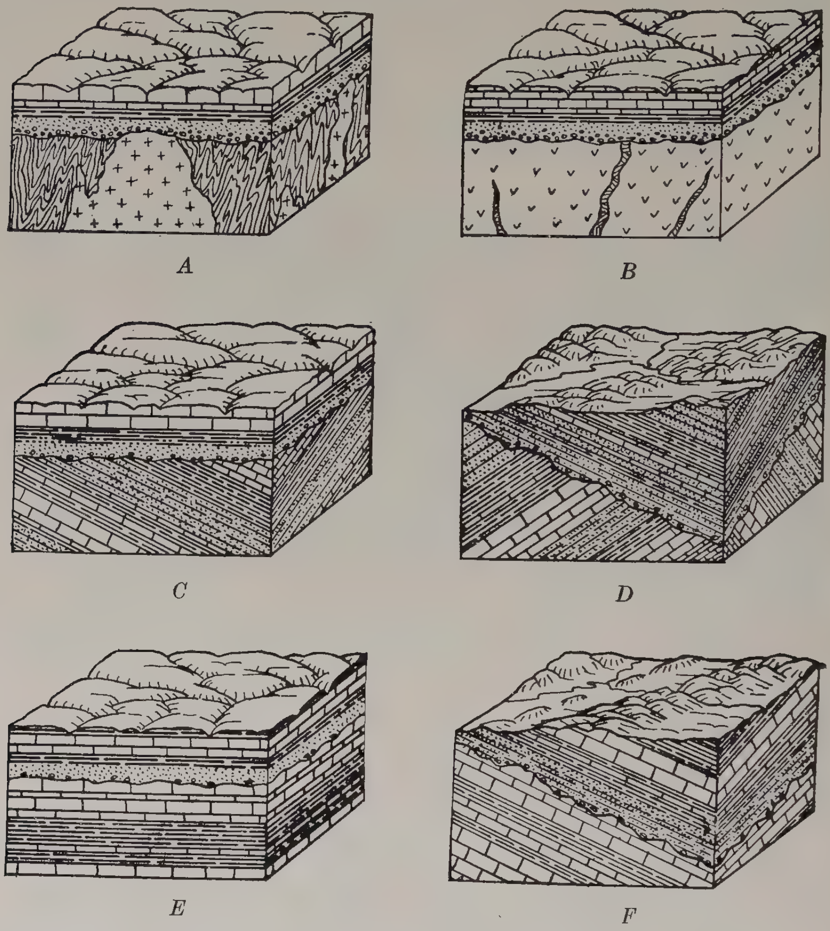


Fig. 229. — Diagrams to illustrate various kinds of unconformity. *A*, sedimentary strata deposited on metamorphic and igneous rocks; *B*, non-conformity on massive igneous rocks; *C*, angular unconformity between two sedimentary series, the later series undisturbed; *D*, angular unconformity, both series tilted; *E*, disconformity, both series of strata horizontal; *F*, disconformity, both series tilted together.

The old and the new formations will then have their stratification planes actually, or practically, parallel. This constitutes an unconformity of the second class, and as it is desirable that it should be distinguished from the other, it has been termed a *disconformity* (Fig. 229, *E* and *F*).

We have then the following cases of unconformity:

Unconformity.
1. Nonconformity, two formations with visibly different structure.
    - *a.* Lower formation of rocks nonstratified, or apparently so.
    - *b.* Lower formation of stratified rocks, tilted with relation to overlying strata (angular unconformity).
2. Disconformity, two formations in parallel position separated by erosion surface.

Obviously the subject of unconformity is closely connected with the study of sedimentary rocks; but a full appreciation of its meaning requires some understanding of crustal disturbance. Strong angular unconformities record severe deformation, either by folding or by faulting; and widespread disconformities indicate warping movements that have involved large areas. A study of unconformities emphasizes the close relationship between crustal movements, erosion, and sedimentation. These ancient records, like the present wasting surface of the land, represent the continual struggle between deep-seated forces that produce irregularity, and surface processes that strive to keep the lands featureless and low.

CHAPTER XIII

EARTHQUAKES

It is difficult to think of earthquakes apart from their relation to human affairs. From the earliest recorded times the recurrent shaking of the solid ground, with consequent destruction on the surface, has been a cause of terror to man. Repeatedly, and in widely separated localities, populous communities have suffered great loss of life and property. Destructive earthquakes recorded during the brief span of human history are numbered by thousands. Geologic evidence indicates that violent shocks have been recurrent throughout the history of the Earth; and there is every reason to expect their frequent occurrence in the future.

The serious aspect of earthquakes from the human viewpoint is realized on review of some major catastrophes. September 1, 1923, approximately 100,000 lives were lost as a result of the Tokyo earthquake, and the estimated property loss exceeded $4,000,000,000. The shocks at Messina in 1908 and at Kansu, China, in 1920 were equally disastrous to life. More than a million and a half persons were killed by ten Chinese shocks between the eleventh and twentieth centuries; and Mallet, a profound student of *seismology* (from *seismos*, an earthquake) estimated that for the whole Earth at least 13,000,000 lives were lost through earthquakes in the course of 4000 years. Some activities of man himself, such as wars, or the operation of automobiles, result in a much higher death rate. Yet earthquakes are especially productive of fear, probably in part because they come without warning, and in part because their cause is more or less mysterious. Study of earth shocks from a geologic standpoint has dispelled a part of this mystery.

Cause of Earthquakes. — An earthquake is a trembling, or undulatory motion, in the more or less elastic rock shell of the Earth, communicated to it by an impulse or shock of some kind just as a bell is set in vibration by a smart tap on its side. The shock or impulse is evidently the immediate cause of the earthquake; but what is the origin of such shocks? Ancient philosophers who sought to explain natural phenomena by natural causes generally connected earthquakes in some way with the weather. Aristotle and Lucretius thought they were produced by winds rushing out of the Earth and leaving voids, with consequent collapse. Modern scientific evidence shows that shocks may arise from several causes, most of which must be considered of minor importance compared with one major source, which appears to give rise to all great earthquakes.

One minor cause is in violent volcanic outbursts, like that of Krakatoa in 1883 and of Bandaisan in Japan in 1888; but earthquakes produced in this way are light in intensity and quite limited in extent. Moreover, many outbursts are not attended by any shocks, or at best by only feeble tremblings, such as occurred during the eruption of Mont Pelée in 1902. Instruments near Kilauea, in Hawaii, record numerous minor tremors — sometimes hundreds of them in a single month. Very few of these are accompanied by visible volcanic activity, although it is probable that shifting of magma at some depth is the principal cause of the local shocks.

Another minor cause of earthquakes may be the sudden caving in of subterranean cavities, or collapse of their roofs under the weight of superincumbent rock masses. This is most likely to happen in regions underlain by limestone, since large quantities of this rock are removed in solution by underground waters.

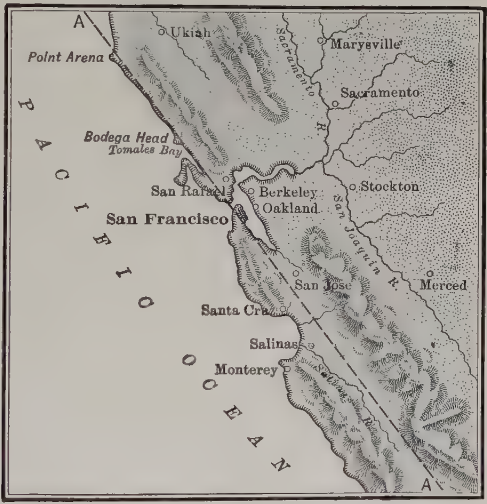


Fig. 230. — Map of a part of California, showing the position and extent of the fault line, A–A, movement along which produced the earthquake of April 18, 1906.

It has now been rather definitely settled that most of the major earthquakes result from the jar given by sudden yielding to strain in the Earth's crust. Such yielding may be by formation of a new fracture, or by abrupt displacement along the walls of an already existent fault. In many areas visited by disastrous shocks the surface of the ground has been broken along fault lines and the amount of displacement is clearly indicated. Commonly these movements take place along old fault zones which bear the marks of repeated displacement. In California a great fracture zone can be followed almost continuously, by means of its peculiar surface expression, from the southern part of the state northwestward for 600 miles. This feature, known as the

San Andreas Rift, passes near the city of San Francisco (Fig. 230). On April 18, 1906, abrupt movement along at least 270 miles of this fracture caused a destructive earthquake.

A careful study along the San Andreas Rift, after the rupture in 1906, yielded valuable information on the nature and amount of the displacement. In general no scarp was made by the faulting, because the motion was almost entirely horizontal, parallel to the fault (Fig. 231). This fact was established beyond question by the offsetting of roads, fences, and other features that extended across the break. Some roads were cut cleanly across, and offset considerably more than their width (Fig. 232). The largest measured displacement was 21 feet. More commonly a part of the motion on a break of this kind is vertical and results in a steep scarp (Fig. 233).

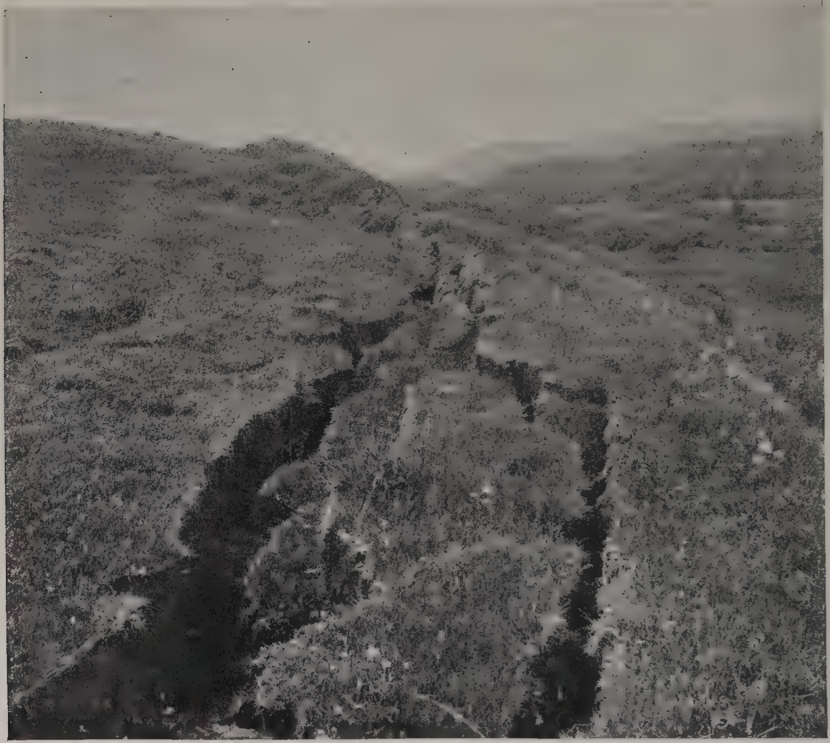


Fig. 231. — Trace of the fault concerned in the California earthquake of 1906. The deep soil above the bedrock broke irregularly. As movement was horizontal, no scarp was formed at this locality. (U. S. Geol. Surv.)

If it is considered that the walls of a fault are pressed closely together, that movement is possible only by overcoming great frictional resistance,

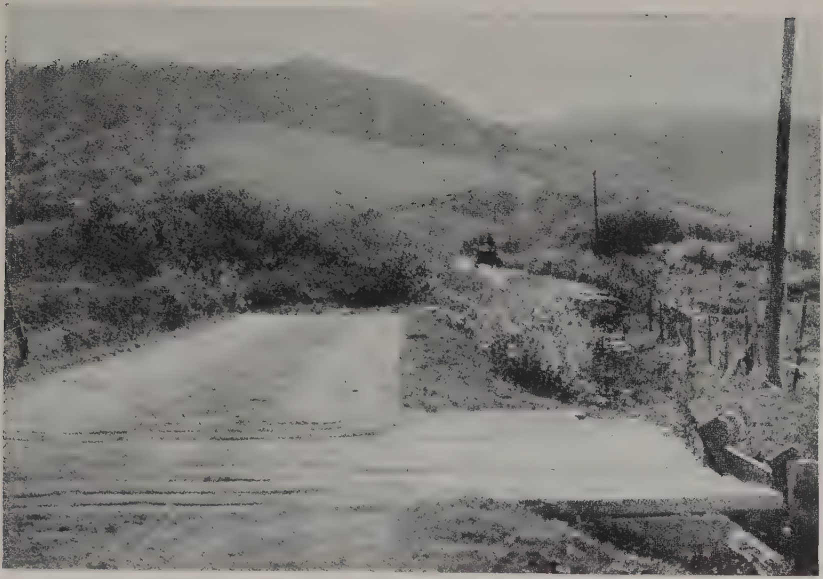


Fig. 232. — Horizontal displacement of a road by movement on the San Andreas Rift in 1906. Before the earthquake the two offset portions of the road were in a straight line. The fault extends from left to right, directly across the road. (U. S. Geol. Surv.)

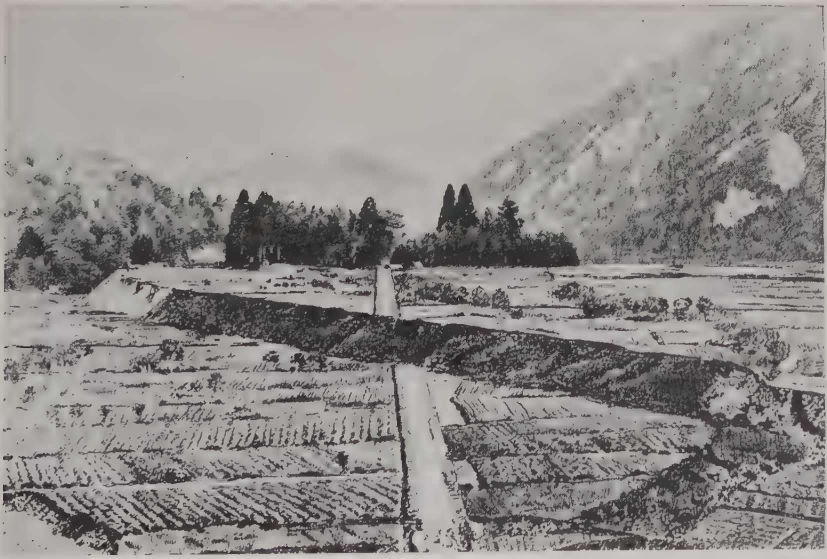


Fig. 233. — Displacement on a fault at Midori in the Neo Valley, Japan, in 1891. The former plain was broken, and one part dropped with relation to the other. Note vertical as well as horizontal offsetting of the road. (I. Ogawa.)

and that the displacement, once it occurs, takes place almost instantaneously, it is not surprising that powerful vibration is set up in the vicinity of the fault line. The exact nature of movement along the San Andreas Rift was made the subject of special study, and it is concluded that mass movement in the crust on opposite sides of the fault was in slow progress for years before 1906. Deformation in the rock was by bending, until the strain could be borne no longer and relief occurred by abrupt slipping along the old fracture. The jar, resulting from the release of energy that had been accumulating for decades, was in effect a heavy blow which made the Earth tremble.

It is not to be supposed that a visible fault appears in every area visited by an earthquake. Commonly the direct evidence of crustal movement is wanting, especially in connection with mild or moderate shocks. It is believed that actual displacement occurs very frequently at considerable depth and does not reach to the surface. Earthquakes of the first rank, however, are in a general way restricted to regions of active faulting, for which there is evidence at the surface. Some destructive shocks originate under the sea; but even in this event it may be evident from soundings that faulting has displaced the sea floor. For example after the Tokyo earthquake of 1923 it was found that an area in Sagami Bay had dropped considerably below its former depth.

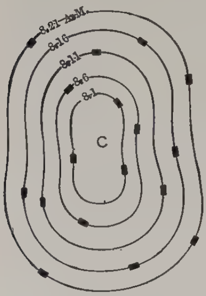


Fig. 234. — Map of coseismal lines. Black rectangles represent points at which the time of the earthquake was recorded accurately.

It was thought at one time that earthquakes were generated from a point at some depth below the surface, and this was called the *focal point*, or *centrum*. The point immediately over this on the surface was called the *epicenter*. This latter point was determined by drawing concentric closed curves, called *coseismal* lines, on a map of the region through points of simultaneous arrival of the waves, as indicated by clocks (Fig. 234). From other data the distance below the epicenter of the focal point was calculated. These methods led to discordant results for some earthquakes, and eventually to the discovery that for any one earthquake there might be several epicenters situated in a line, or that where earthquakes habitually occurred in a given region the different epicenters were situated along a line. Such a line probably represents a fault, even if there is no surface evidence of its existence. The terms centrum and epicenter still have value, although it should be understood that they are not points.

Effect of Shock. — It is important that we distinguish clearly between cause and effect in earthquake phenomena. The displacements shown in Figs. 232 and 233 are not the *results* of earthquakes as is commonly supposed; they represent the *causes*. The effect of the sudden movement along a fault is to set up vibrations that move outward from that place, and these constitute the earthquake, as it is perceived at a distance. Thus the earthquake is propagated as a series of waves in the highly elastic body of the Earth. When these elastic waves emerge at the surface the loose ground is thrown into rapid vibration, ordinarily with an amplitude not exceeding a few inches. In bedrock the amplitude is only a fraction of an inch. The actual amount of movement reaches a maximum in deep alluvium that is saturated with water. On terrane of this kind destructive effects are greatest. For example at San Francisco the devastation was most acute on the low alluvial flat near the bay. Buildings on solid rock, even much nearer the fault, suffered less damage. The contrast in behavior of the deep alluvium and of bedrock may be illustrated by striking sharply a bowl in which there is jelly. The bowl is set vibrating with resulting sound, but actual motion in the walls of the vessel is imperceptible. However the same impulse transmitted to the jelly sets up longer waves that are visible. Excessive destruction on wet alluvium is caused by this larger wave motion as compared with the behavior of solid rock. In any kind of material the motion is very complex (Fig. 235).

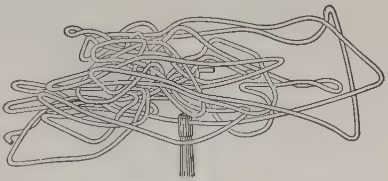


Fig. 235. — Wire model showing path traveled by a particle of matter during an earthquake; after Sekiya.

Examples. — On August 31, 1886, the city of Charleston, South Carolina, was visited by a severe earthquake which did great damage. The shock was distinctly felt as far away as Chicago, a distance of 800 miles. This shock is of special interest, because from general considerations it does not appear that severe crustal disturbance should be expected at Charleston.

In 1899 a great earthquake took place in southern Alaska. As the region is mostly uninhabited the shock passed almost without notice at the time. Studies which have since been made show that considerable alterations in topography took place at the time of its occurrence, especially about Yakutat Bay. Marked changes were also induced in the great glaciers of this region by the shattering of the ice and by snowslides from the mountains.

In August, 1906, the coast of Chile was visited by a severe earthquake, which did great damage in Valparaiso and other places. After-shocks continued for a long time while readjustment along the fault was going on. The west coast of South America is noted for its earthquakes, in connection with which notable elevation of the coast line has occurred.

One of the greatest disasters in modern times occurred on Dec. 28, 1908, when Messina and Reggio, cities on the narrow strait which separates Sicily from the mainland of Italy, were completely destroyed by a terrific shock. Evidently the area is in a zone of crustal weakness and readjustment, as severe earthquakes have occurred repeatedly.

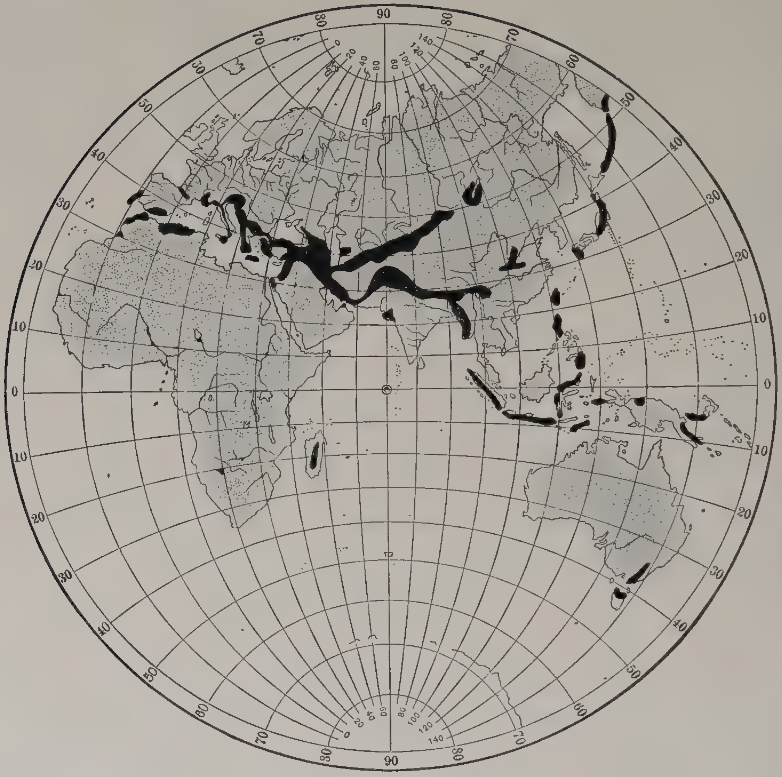


Fig. 236. — Map of seismic belts in the Eastern Hemisphere. On S. L. Penfield's stereographic projection. (Compare map of volcanic belts, Fig. 170.)

The great Tokyo earthquake of 1923 is remarkable for the very large changes in topography that were produced by the crustal disturbance. Parts of the shore around Sagami Bay were lifted up as much as 6 feet. In addition to large depressions on the floor of this bay, other parts of the floor were lifted several hundred feet, making shoals where there had been deep water.

These are only a few examples out of many that might be selected. Scarcely a day passes that shocks are not recorded from some part of the world by earthquake observatories.

Seismic Belts. — Although earthquakes occur in all parts of the world, they are most likely to happen in certain well-defined tracts, which lie in the two great seismic belts. One of these follows the western coast of North and South America, the Aleutian Islands, and the island groups along the eastern coast of Asia, and thus borders the Pacific Ocean on the east, north, and west. The other includes the Mediterranean, the Alps, the Caucasus, the Himalayas, and continues into the East Indies, where it intersects the first belt at a large angle. (Figs. 236 and 237.) In a general way these zones coincide with the great volcanic belts (page 225); and this fact might appear to support the idea that volcanoes are an important cause of earthquakes. However, since the belts correspond

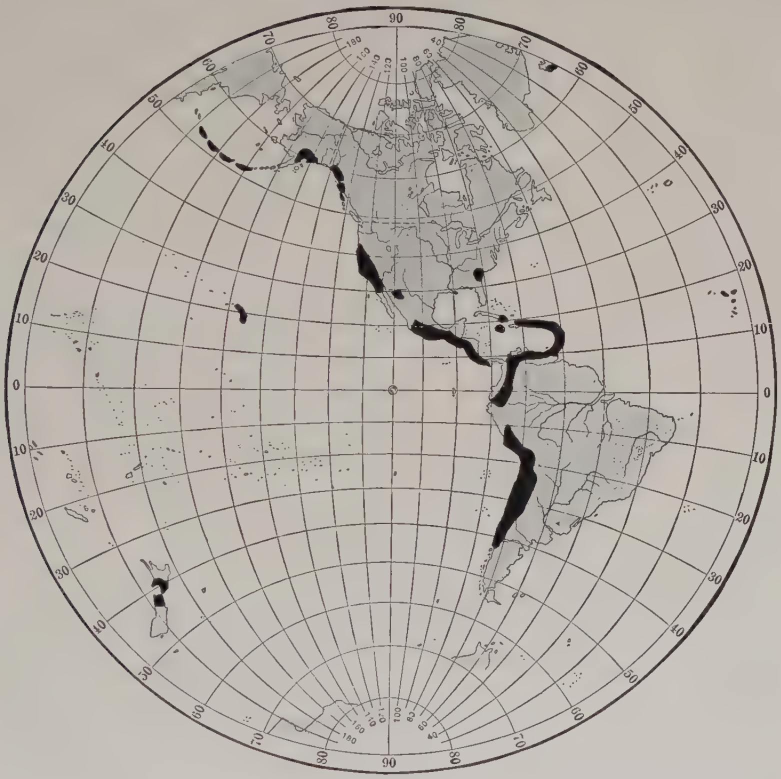


Fig. 237. — Map of seismic belts in the Western Hemisphere. On S. L. Penfield's stereographic projection. (Compare Fig. 171.)

closely to young mountain systems and other marks of recent crustal movement, it is probable that both earthquakes and volcanoes have a common cause in this disturbance of the lithosphere. It is a notable fact that where the seismic belts lie directly along the continental borders, as on the coast of Chile and the eastern coast of Japan, the land descends sharply, without any broad intervening shelf, to great depths of the ocean. Some of these steep slopes descend into *foredeeps,* which are great troughs that appear to be sinking, while the bordering lands are rising. We conclude that these are zones of weakness in the Earth's crust where strains are being constantly relieved by movements, and in which, therefore, earthquakes are continually recurring.

It is commonly thought that certain regions are practically exempt from danger of earthquakes because no real disaster has happened in them since they have been settled and cities have sprung up. It is true that most of the Atlantic coasts, and large areas in continental interiors, are relatively free from earthquakes. However, the experiences of New Madrid in 1811 and of Charleston in 1886 are a warning that no locality may be entirely exempt. Even in New England, which is not recognized as a seismic tract, there has been an average of one perceptible tremor a year since the settlement of the country. Probably none of the shocks has been of maximum intensity, although several have caused some destruction.

Submarine Earthquakes; Tsunamis. — The location of seismic belts suggests that many earthquakes originate under the ocean. Their occurrence beneath the sea is shown by shocks communicated to vessels on the surface above, and by rupturing of submarine cables. Since the invention of sensitive instruments by which it is now possible to record distant earthquakes and determine their location, it has been learned that a large proportion of all earthquakes occur on the floor of the Pacific. The most conspicuous mark of a submarine earthquake is the

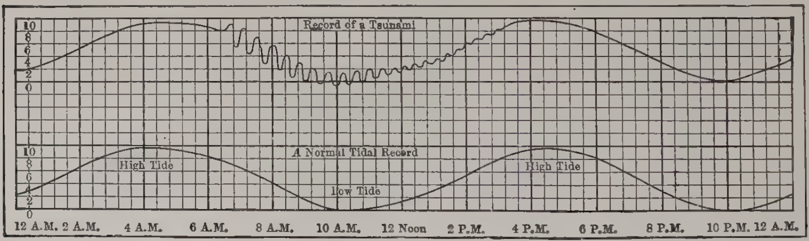


Fig. 238. — Record of tsunami by tidal gauge (upper curve). Vertical lines represent time spacing on the paper, driven horizontally by clockwork. Horizontal lines show height in feet as recorded by the rising and falling pencil of the gauge.

huge wave that commonly is generated in the ocean by disturbance of the floor. Such waves have long been known as tidal waves, a misleading name since they have no connection with the tide. They are now generally known to seismologists by their Japanese name *tsunamis;* or they may be called seismic sea waves. Some of are immense size, measuring 100 or even 200 miles from crest to crest, and as much as 40 feet high. They are so broad that in the open sea they are not ordinarily perceived; but on approaching the coast they may pile up in huge breakers and, sweeping far inland, cause enormous damage and loss of life.

Lisbon in 1755, Japan in 1854 and in 1896, Peru in 1868, suffered from great and disastrous tsunamis. The number of victims of a single inundation of this kind has been as great as 20,000. These vast waves are felt over whole oceans and move with tremendous speed, — from 300 to 500 miles per hour. Those from Japan have crossed the Pacific in about 12 hours. At such distances their height may be only a few inches; but the ebb and flow of from 15 to 30 minutes, like small subordinate tides, are registered as wavy lines on the record of a tidal gauge (Fig. 238).

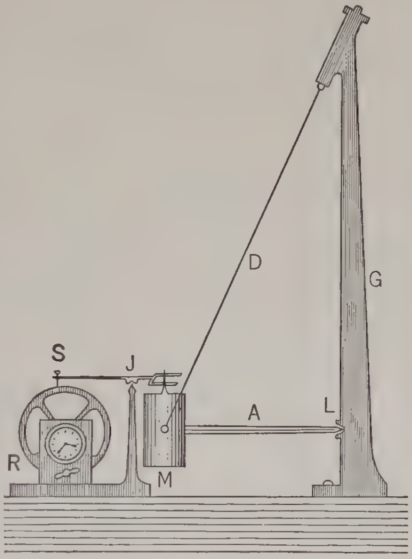


Fig. 239. — Diagrammatic representation of a seismograph. The upright post (G) is attached firmly to bedrock. The heavy weight (above M) is connected with the post only by a freely-moving joint (L) and a flexible wire (D). Records are made by the stylus (S, J) on the revolving drum (R), which is driven by clockwork. (Sieberg.)

Recording Earthquakes. — Very delicate instruments, called *seismographs* (Fig. 239), record the tremors due to distant earthquakes; and the study of these records has led to important geological conclusions. The principle upon which the commonest instruments are constructed is simple. If a heavy mass of metal be suspended like a pendulum, owing to its inertia it will remain for a time at rest when the shock arrives, while the bedrock vibrates beneath it. A pencil of some kind is secured to the suspended weight, and rests lightly on a paper or other medium suitably prepared to record the motions of the pencil. When the bedrock oscillates the heavy weight acts essentially as a steady point; but the vibration is transmitted to the pencil, and may be magnified as many times as desired by a simple mechanical device. If the paper, instead of being made fast, be a strip continuously carried along by clockwork, the pencil when at rest will draw a straight line upon it; when vibrations of the Earth occur the line will bend sinuously from one side to the other (Fig. 240). Such a record is known as a *seismogram*. Some instruments have, instead of a point or pencil, a small mirror that throws a beam of light upon a sheet of photographic paper. The seismogram is revealed only after the sensitized paper is developed.

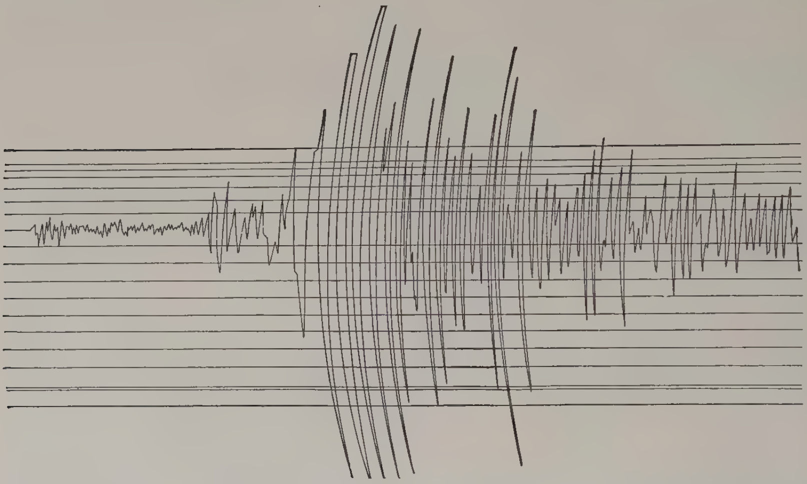


Fig. 240. — Record of the earthquake in Messina on Dec. 28, 1908, as shown by a seismograph in Göttingen, Germany, over 1000 miles distant. The actual vibration experienced by the instrument is greatly exaggerated in the seismogram.

Since the intervals of time are marked on the moving paper, the instrument records the time of arrival of the shock and also the duration. The directions of diversion of the markers from their regular paths show also the direction from which the shock has come.

Seismograms. — The study of seismograms of distant earthquakes has led to the discovery that the main shock is preceded by smaller rapid vibrations known as the preliminary tremors. Thus a normal seismogram has the characters seen in Fig. 241. It has been determined that these preliminary tremors represent elastic impulses that come by the shortest path *through the Earth*, that is, in the general direction of a chord from the seat of disturbance to the recording station; whereas the later large vibrations represent those elastic waves that have traveled by a longer route over the surface circumference. The first preliminary tremor (*ab*, Fig. 241) is caused by a compressional or longitudinal wave (commonly known as the *primary*), which travels several miles per second; the other preliminary (*bc*, Fig. 241) represents a transverse wave motion (the *secondary* wave), which travels at a distinctly slower rate. Therefore the time interval between the two preliminary tremors is proportional to the distance traversed, and from this information the distance between the seat of the shock and the seismograph can be calculated accurately. If distances from at least three separate stations are computed, and circles are drawn on a map with these distances as radii, the circles will intersect in a common point, which is the locus of the earthquake.

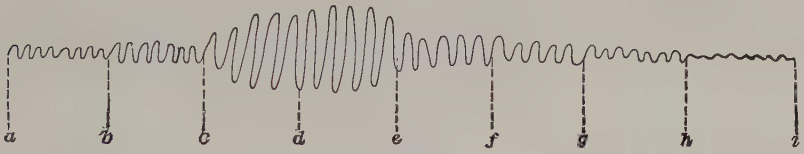


Fig. 241. — Seismogram of distant earthquake; *ab*, first preliminary tremors; *bc*, second preliminary tremors; *ce*, main shock; *fh*, later phases; *hi*, tail. (After Omori.)

Geological Deductions from Seismograms. — The fact that the preliminary tremors, which travel through the Earth, arrive at distant points so long a time ahead of the main shock, cannot be explained alone by the shorter path traveled. The time interval shows that they are also propagated at a much greater rate of speed than the vibrations traveling in the outer shell of the Earth. The deduction from this is that they move in a more elastic medium than the superficial part of the crust. Moreover, the concordant results in different directions show that inside of the outermost layer, which we know is heterogeneous in composition, the Earth is homogeneous, or regularly arranged around its center in structure; or, if nonhomogeneous, the heterogeneous parts are relatively so small and numerous that different paths of considerable length through them give the effect of uniformity. Furthermore the average velocity increases with the distance of the recording station; thus the average rate of transmission along *sz*, Fig. 242, is greater than along *sy*, which in turn is greater than along *sx*. Velocities of the primary and sec-

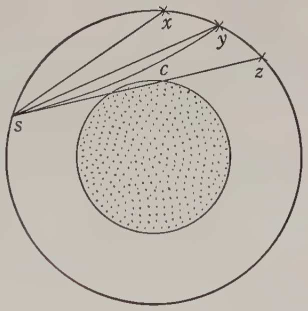


Fig. 242. — Paths of transmission of earthquake shock through the Earth. *s*, seat of shock; *x*, *y*, *z*, recording stations. Stippled area represents the central core, which apparently has a composition or state different from that of the outer zone. Stations *x* and *y* receive a complete record of the shock at *s*; but the secondary waves do not reach *z*.

ondary waves, as calculated from available earthquake data, vary with depth as shown in the following table:

Depth Below Surface in Miles	Velocity, in Miles per Second	
	Primary Wave	Secondary Wave
0	4.46	2.47
228	5.43	3.03
784	7.39	4.14
1680	7.95	4.25

These results show not only that velocity increases down to a certain depth, but more and more slowly as the depth increases, and this would seem to indicate that the density and elasticity of the Earth increase with depth down to a certain region. It is a matter of considerable importance that the rate of propagation actually diminishes below a depth of nearly 2000 miles, and the secondary wave dies out entirely. Seismographs whose distance from the locus of an earthquake exceeds one-third the Earth's circumference (120°) receive no record of the transverse wave. There is strong indication, therefore, that the Earth has an inner core, about 4000 miles in diameter, whose composition or state is different from that in the shallower zones. There is good reason for believing that the core consists chiefly of metal instead of rock. Some scientists have suggested also that this metallic core may be in the fluid state. This would explain the failure of the transverse waves to penetrate the core, as such waves are transmitted only through elastic solids.

Geological Effects of Earthquakes. — There are several geological effects from earthquakes, but they are, comparatively speaking, of minor importance. The loose mantle of soil and other débris is often ruptured by the passage of the wave with the formation of fissures, which may be of some depth. A more important effect is the starting of landslides and avalanches in mountainous regions, through the jarring of the Earth. A variation in the flow of water from springs, or even the forming of new springs, has also been observed.

Much more important are the movements of the crustal blocks at the time of earthquakes; but as previously emphasized, these are the cause and not the effect of the shocks.

CHAPTER XIV

METAMORPHISM AND METAMORPHIC ROCKS

Definition of Metamorphism. — In addition to the igneous and sedimentary rocks previously described there is a third class termed the metamorphic, whose distinctive characters are due to metamorphism. These rocks are the records of the remarkable transforming power of certain geologic forces that are at work within the Earth's crust.

Metamorphic means *changed in form*, and *metamorphism* is a general term for all those changes by which the original characters of rocks are more or less thoroughly altered, so that the component minerals or textures of the rocks are transformed into new minerals or textures, or both. Some metamorphic rocks pass gradually into those whose fossils and stratification prove them to be undeniably of sedimentary origin, whereas other metamorphic rocks grade into rocks whose characters show conclusively that they are of igneous origin. From these transitions we learn that some metamorphic rocks have been formed from sedimentary and some from igneous rocks. As we shall see later, some metamorphic rocks have also been derived from other metamorphic rocks.

Some metamorphic changes are so profound that the resultant product no longer resembles the rock from which it was derived but has become a new rock. Sedimentary rocks thus thoroughly *metamorphosed* are more coarsely crystalline, and the fossils that they may have contained and even the marks of stratification have been completely obliterated. Igneous rocks also, if severely metamorphosed, have lost their original distinctive features.

Limestone, for example, may be metamorphosed into coarse-grained marble with consequent loss of color and obliteration of the fossils that were in it; basalt may be converted into a green, slaty rock that gives no hint of its original igneous nature. Rocks that represent the stages of transition between the limestone and the marble or between the basalt and the green slate can be found; but under metamorphic rocks we include only those which have been so profoundly changed that their original outward characters have been either entirely obliterated, or nearly so, and distinctly new rocks have been formed. We say that it is the " outward " characters which are obliterated, for the bulk chemical composition of a rock as a rule remains unchanged during metamorphism. The various changes that rocks undergo from the effects of *weathering* might in the strict etymologic sense of the term be classed as metamorphic. But they have been already discussed under the work of the atmosphere and the production of soils; therefore these agencies and the weathered rocks and soils that result from their action are not considered in this place.

KINDS OF METAMORPHISM

Three kinds of metamorphism are recognized: 1, contact metamorphism; 2, dynamic metamorphism; and 3, load metamorphism. Contact metamorphism is produced by the action of intrusive igneous masses on the rocks into which they were intruded. The effects produced are due chiefly to the heat supplied by the magma and the hot gases that issued from it when it consolidated. These effects are of course limited to the vicinity of the contact with the igneous mass, and hence the term contact metamorphism; but the term igneous metamorphism, by emphasizing the agency that produces the changes, is more appropriate. The most obvious form of contact metamorphism, probably the most impressive to the layman because coming more nearly within the ken of ordinary experience, is the conversion of a coal bed into a layer of coke by an injected sill.

Dynamic metamorphism results from the action of tangential pressure in the Earth's crust, as displayed in folding of the strata and in great overthrust faults. As a metamorphism of this kind often accompanies profound dislocations of the crust, it is called by Heim, the great master of metamorphic geology, *dislocation-metamorphism.*

In places rocks appear to have been metamorphosed without either the intervention of igneous masses or of dynamic metamorphism. Here the cause of the metamorphism was apparently that the rocks were formerly deeply buried, having become depressed in the crust under a heavy load of overlying strata. The increase in temperature brought about by the deep subsidence has caused the development of new minerals and textures, and the heavy pressure due to the load favored the production of heavy minerals, such as garnet, which have the minimum volumes. As the Earth's own heat is the main factor here in bringing about the transformations, the process is sometimes called geothermal metamorphism.

The best established example of metamorphism of this kind is furnished by the German potassium-salt deposits. These salts were laid down in an evaporating arm of an ancient sea; they were many and of complex compositions; they were stable, however, under the conditions of moderate temperature then prevailing. Subsequently the basin in which they were deposited slowly subsided and they became covered by 20,000 feet of sedimentary beds. The temperature of the deeply buried salt beds rose to that determined by their depth in the crust, and drastic rearrangements in the composition of the salts took place and many new minerals were produced.

Silicate rocks are far less sensitive than the potassium-salt minerals. Nevertheless, certain formations of pre-Cambrian age are believed to have acquired their metamorphic condition by load metamorphism.

CONTACT-METAMORPHIC ROCKS

As previously explained, the term contact metamorphism is used to denote the changes that are induced in rocks by the intrusion into them of a mass of magma.

The most noticeable effect of contact metamorphism is a baking, hardening, or toughening of the intruded rocks in a zone that surrounds the intrusive igneous mass.

Width of Contact Zone. — The width of the contact-metamorphic zone depends chiefly on the size of the igneous mass. The widest zones occur around stocks and batholiths. Around them the contact zone varies in width from some hundreds of yards to a mile or more; but adjacent to a small intrusion such as a dike, it may be only a few feet. Lava flows produce at most a slight baking of the soils or rocks on which they rest.

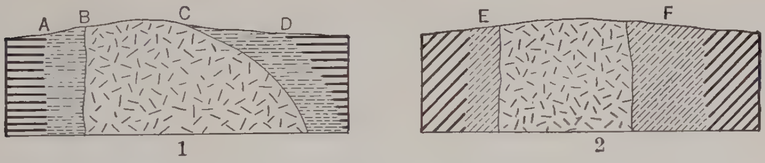


Fig. 243. — Sections of intruded stocks and their contact zones. In 1, the breadth on the surface CD is greater than AB, depending on shape of igneous mass. In 2, the width F is greater than in E, because the inclined beds allow the rising hot vapors from the igneous mass to penetrate farther on this side than on the other.

The width of the contact zone around one and the same igneous mass varies, for it is controlled by the configuration of the igneous body and by the attitude of the surrounding rocks. Thus in Fig. 243, section 1, as a result of the lesser slope of the contact a wide zone is produced at *CD*, much wider than that adjacent to the vertical contact at *AB*. And in section 2 the beds at *F*, which slope toward the igneous rock, tend to have their bedding planes opened, and to furnish easy passageways to the emanations from the cooling magma, which are the chief agents in carrying the heat and producing the metamorphism.

Contact Metamorphism of Rocks of Different Kinds. — The extent and the intensity of contact metamorphism depend very much, in addition to the factors already discussed, on the kinds of rocks surrounding the igneous mass. For our purpose here the sedimentary rocks may be divided into the three groups: *sandstones*, *shales* (and clays), and *limestones.* On pure sandstone the effect is rather small, though near the contact the sandstone may be changed into quartzite — a compact rock so firmly cemented that it fractures across the grains, instead of around them. Limestone is changed into marble, the masses of which may extend for considerable distances from their contact with the igneous rocks. Shales show the most notable and, generally, far-reaching results. The soft shales are greatly hardened, and near the contact they are converted into a rock known as *hornfels*, which to the unaided eye strongly resembles a black-fine grained igneous rock, such as basalt.

The most interesting results are produced in limestones, especially impure varieties. Not only are the limestones turned into marble, but a great variety of minerals are newly formed in them. Thus when the limestone is heated above 500° C. the silica tends to drive out carbon dioxide, and calcite is changed into calcium silicate. If the limestone contains dolomite, a pyroxene is formed.

If other impurities occur in the limestone, such as clay furnishing alumina, and iron oxides, many other new minerals will be formed. These more complex reactions may be illustrated by the following example:

Calcite + Clay + Quartz = Garnet + Carbon dioxide + Water.

Thus a limestone containing clay and sand as impurities may be changed into garnet with evolution of carbon dioxide and water.

In the normal contact-metamorphic zone the effects produced are entirely owing to the heat given off by the igneous mass: the surrounding rocks become heated to a high temperature and new minerals are produced by the recombination of the elements already present in them, and consequently the chemical composition of the rocks remains unchanged. A marble, for example, has the same chemical composition as the limestone from which it was formed. If, however, the igneous mass during its consolidation gives off gases carrying iron, silicon, boron, etc., vastly different results are produced in the surrounding rocks. The most striking results are produced in limestones, because of the readiness with which limestones react with the magmatic gases. Many new minerals are formed, often beautifully crystallized. The resultant contact-metamorphic rock differs greatly in composition from the original rock: substances have been added in large amount by the gases that streamed from the magma. If the gases contained iron, copper, or tungsten in notable quantity, valuable ore deposits may be formed in the limestones as incidental by-products of metamorphism of this kind.

"Crystalline Schists"

Most metamorphic rocks have *not* been formed by contact metamorphism, at least under the static conditions that prevailed while the rocks of the hornfels type were produced. So well known is this fact that the term metamorphic rocks when used without qualification referes to the great group that is not of contact-metamorphic origin. We lack a good distinctive designation for this group, and for want of a better name they are called the *crystalline schists*, in virtue of two characteristic features that most of them have: their obviously crystalline appearance and their foliated or schistose texture.

Many of these crystalline schists are the products of dynamic metamorphism, during which the rocks from which they were derived were forced by heavy differential pressure to flow in the solid state. As a result of this flowage there were developed either new textures, or new minerals, or both. Examples of such rock flowage are well shown in the Alps, perhaps the most marvelously folded tract in the world, where the limbs of the folds have become attenuated to one-tenth, even to one-hundredth of their original thickness and the arches of the folds have become enormously thickened: manifestly solid material has flowed from the flanks of the folds to the points of flexure.

Foliation. — Most metamorphic rocks are visibly crystalline to the unaided eye, and they have a parallel arrangement of their constituent minerals. This parallel arrangement gives the rock a foliated texture (from *folium*, a leaf), and a rock having it is known as a foliate. By reason of it a rock tends to split, or cleave, more or less perfectly into flakes or slabs parallel to the foliation. A coarsely foliated rock is called a gneiss, and one in which the foliation is well developed and closely spaced is termed a schist. Although foliation is the characteristic texture of metamorphic rocks, there are a few, such as serpentine, marble and quartzite, which commonly show no trace of this texture. In some rocks the parallel texture is straight or nearly so for considerable distances, as seen in Fig. 244; commonly, hoewver, the banding is very much contorted, bent, or curled, showing the amount of deformation to which the original rocks were subjected (Fig. 247).

The foliated texture is due to the occurrence of the component minerals in separate layers or flat lenses, or to the parallel arrangement of prismatic or tabular minerals, such as hornblende or mica, or to a com-

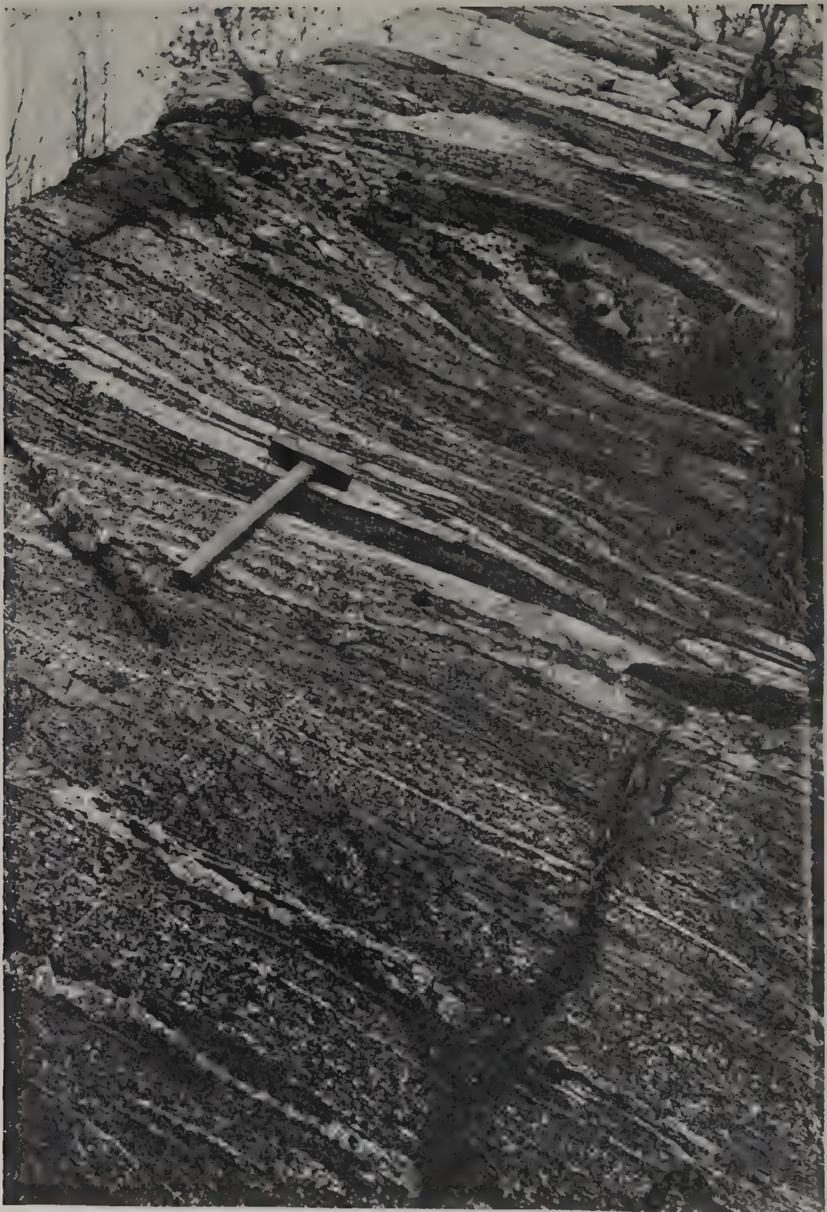


Fig. 244. — Banded gneiss, Portland Township, Ottawa Co., Quebec. (Geol. Surv. of Canada.)

bination of both modes of arrangement. It is a result of the granulation and recrystallization to which the original rocks were subjected, and has been imposed on igneous and sedimentary rocks alike. The resemblance of the banding or lamination of schistose rocks to stratification led in the past to the erroneous view that all of them were derived from stratified rocks; that some have also been made from igneous rocks was learned much later.

Slaty Cleavage. — The cleavage exhibited by metamorphic rocks is most remarkably developed in slates, so characteristically that this variety of it is called *slaty cleavage.* Slates used for roofing, blackboards, and other purposes show this cleavage. The question of its origin has stimulated much investigation, along both experimental and mathematical as well as geological lines. From these studies it has become clear that the cleavage is the result of great pressure on the material, and that *the plane of cleavage is at right angles to the direction of pressure that produced it.* When fine-grained sediments, such as muds and clays, are subjected to intense pressure, the oblong particles in them tend to rotate so that their lengths are perpendicular to the direction of pressure; they also tend to become flattened perpendicularly to it. Important also is the fact that many of the platy or elongated minerals, such as mica and chlorite, have an excellent cleavage parallel to the flat directions. A considerable part of the minerals in the slate, such as the micas, were not originally present in the sediments, but were formed during metamorphism accompanying the pressure, and as they grew they set themselves in parallel orientation. All these features tend to give the rock a capacity to cleave readily in one direction.

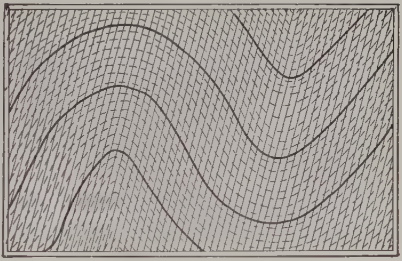


Fig. 245. — Slaty cleavage (represented by the nearly vertical lining) in folded beds.

The cleavage planes do not necessarily bear any fixed relation to the bedding planes. The beds were laid down in horizontal position and the direction of pressure is also horizontal or nearly so. Therefore the cleavage planes, being developed at right angles to this pressure, commonly cut the bedding at highly inclined or right angles. For the beds may become folded before the pressure becomes intense, and the cleavage planes, being developed after the folding, will consequently intersect the bedding at various angles, though these cleavage planes themselves are all strictly parallel to one another (Figs. 245, 246, and 248).

Although most slates have been made from fine sediments, such as muds and clays, slaty rocks have also been produced by the shearing of fine-grained igneous rocks, such as felsites and basalts, and beds of volcanic ash. In the making of slates the original characteristic features of the rocks become greatly distorted; thus fossils and pebbles in the stratified rocks and embedded crystals and other structures of the igneous rocks are flattened into lenses or squeezed out into cylinders.

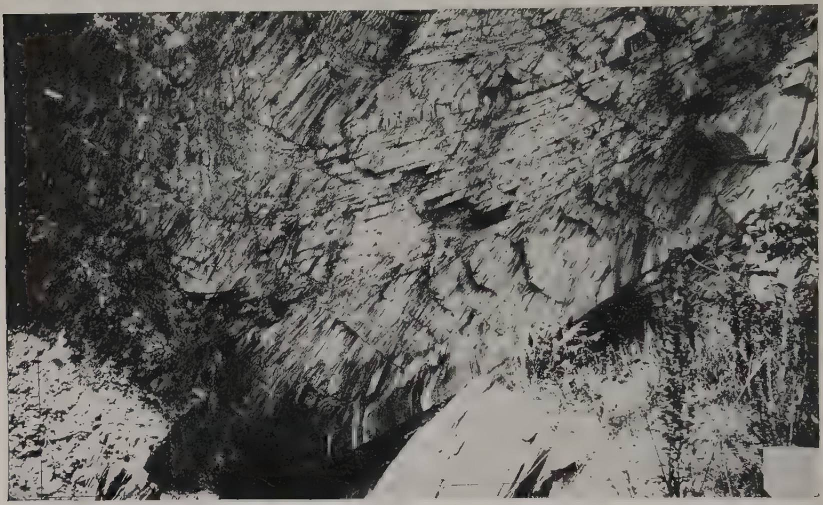


Fig. 246. — Slaty cleavage cutting at a high angle beds folded in a syncline. Slatington, Pa. (U. S. Geol. Surv.)

Places of Occurrence. — Crystalline schists are widely distributed over the Earth's surface, and in some regions they are the only rocks exposed over extensive areas. Metamorphic rocks are particularly abundant in the most ancient (pre-Cambrian) formations the world over. In the United States they occur quite generally in New England, in the Adirondacks, and in a strip of country running from New York to Georgia. New York, Philadelphia, Baltimore, and Washington are built on metamorphic rocks. Metamorphic rocks are well shown in the inner gorge of the Grand Canyon of the Colorado, and in many other places.

The metamorphic rocks form the cores of many mountain ranges, in which they have become exposed by denudation. The structure of these mountain ranges, as will be discussed later, includes folded strata, and the degree of metamorphism of the rocks is proportional to the closeness and intricacy of the folding. In other ranges, however, strata just as closely and intricately folded have not been metamorphosed. Hence mere intricacy of folding does not determine metamorphism. Probably the depth at which the folding takes place within the crust and the speed with which it occurs are the controlling factors.

KINDS OF METAMORPHIC ROCKS

Introductory. — Because the metamorphic rocks are derived from igneous, sedimentary, and also from metamorphic rocks, it follows that there must be an extraordinary diversity of metamorphic products. Yet in the same way that we were able for ordinary purposes to gather the igneous and stratified rocks into a few groups, so we can consider the metamorphic under the few most important types.

Classification. — It is possible to show in a general way the relation between the most common sedimentary rocks and their metamorphic derivatives in the following table:

Sediments	Sedimentary Rocks	Metamorphic Rocks
Gravel	Conglomerate	Gneiss
Sand	Sandstone	Quartzite
Silt and clay	Shale	Slate, phyllite, and various schists
Calcareous deposits	Limestone	Marble

The igneous rocks, it will be recalled, are roughly divided into two main groups, the one chiefly composed of light-colored feldspathic minerals, and the other mostly of dark ferromagnesian minerals. We can illustrate in a rough way the relation between them and their metamorphic derivatives in the following table:

Igneous Rocks	Metamorphic Derivatives
Coarse-grained feldspathic types, such as *granite*, etc.	*Gneiss*
Fine-grained feldspathic types, such as *felsite, tuff*, etc.	*Slate* and *schists*
Ferromagnesian rocks, such as *gabbro* and *basalt*	*Hornblende schists* and *serpentine*

Comparison of the tables will show that gneisses and schists may have diverse origins, as previously pointed out. Combining the results of these tables, we obtain the following main groups of metamorphic rocks, distinguished according to their mineral composition or by their texture, or by a combination of both.

Grouping of Metamorphic Rocks

1. *Gneiss.*
2. *Quartzite.*
3. *Mica schist.*
4. *Slate* and *phyllite.*
5. *Hornblende schist; talc* and *chlorite schists.*
6. *Marble;* mixed carbonate-silicate rocks.
7. *Serpentine.*

Gneiss. — The most common variety of gneiss (pronounced *nice*) consists, like granite, of quartz, feldspar, and mica, but as the mica is arranged in more or less parallel bands, the gneiss has a rude cleavage. Some hornblende and other minerals, such as garnet, also occur, giving different varieties. Gneisses range in color from light to dark, and from fine to coarse in grain. Varieties transitional between granite and gneiss are very common. In those gneisses made from conglomerates the original pebbles may still show as lenticular masses.

Gneiss is one of the most common of metamorphic rocks, and its many varieties have been formed under very diverse conditions: some under conditions of mild dynamic metamorphism in which the minerals have been merely mechanically deformed, and others under the conditions of most intense metamorphism, in which the mineral composition has been wholly reconstituted.

Some granite gneisses are called *primary* gneisses, because the foliated or gneissic texture has been assumed during magmatic flowage or by the shearing of magmas still in a partly consolidated or viscous condition. Batholiths that have been intruded during mountain folding generally have foliated borders of primary gneiss. Figure 247 illustrates a typical exposure of gneiss.

Quartzite. — Quartzite is a rock composed of quartz grains which are so firmly cemented that fracture takes place through the grains, instead of around them. Originally the rock was a sandstone and has been changed to a quartzite either by filling the pore space of the original sandstone through the deposition of quartz from circulating water, or through elimination of the pore space mainly by compaction and rearrangement of the quartz already present in the rock. Although quartzites of this origin are intimately associated with foliated rocks, they themselves rarely show foliation.

Quartzites are generally compact hard rocks of light colors — white, gray, reddish, or buff — and are likely to be of vitreous appearance.

Mica Schist. — Mica schist is the most widely distributed and important member of the great class of crystalline schists. The essential minerals are quartz and mica, and it is especially the mica that gives the rock its distinctive character. Different varieties of mica occur; the most common is a silvery white muscovite, which gives the rock a brilliantly spangled appearance, and the black mica — biotite — also is common. The micas are in irregular leaves or tablets with their cleavage planes oriented in the direction of the schistosity, and it is in fact this parallel arrangement of the micas that produces the extraordinary fissile character of mica schist. Some mica schists are dotted with dark-red garnets in well-formed crystals.

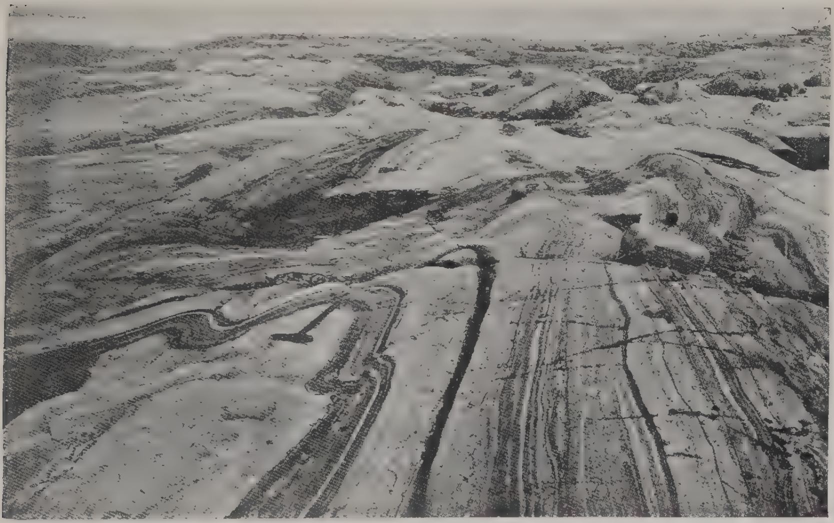


Fig. 247. — Contorted gneiss; Fullerton, Hudson Bay, Canada.
(Geol. Surv. of Canada.)

Mica schists are rocks that have attained a high-grade metamorphic condition. The initial material from which they were derived was an argillaceous sediment — a shale of some kind — and the transformation of the dull, amorphous substance of the shale to a brilliantly spangled mica schist is in a way as remarkable as the metamorphosis of a chrysalis into a beautiful butterfly.

Slates. — The origin of slates from fine-grained sediments, such as muds, clays, and ash deposits, by the action of compressive forces has already been discussed. Although they are of various colors — red, green, gray, etc., — the most common is dark gray to black, due to carbonaceous material in the original muds. As is well known, they are quarried for roofing slates, blackboards, and other purposes (Fig. 248). They are closely related to shales, but the distinction between them is that the cleavage or fracture of a shale is conchoidal, or "shelly" (whence its name, shale), and is parallel to the bedding planes; whereas in slates it is a secondary induced structure, which, as previously stated, commonly cuts across the bedding. Slates are metamorphic rocks of very low grade.

Phyllites. — Phyllites resemble slates, but their constituent mica is coarser, which gives them a silky, glimmering luster. Most of them are transitional between slate and mica schist and are the results of an intermediate grade of metamorphism: more intense than necessary to produce a slate and not sufficiently intense to produce a schist. Some, like the ordinary slates, have been formed from sediments, but are more highly metamorphic. Others have been made from igneous material — felsite lavas, tuffs, etc., — by shearing and accompanying agencies of metamorphism.

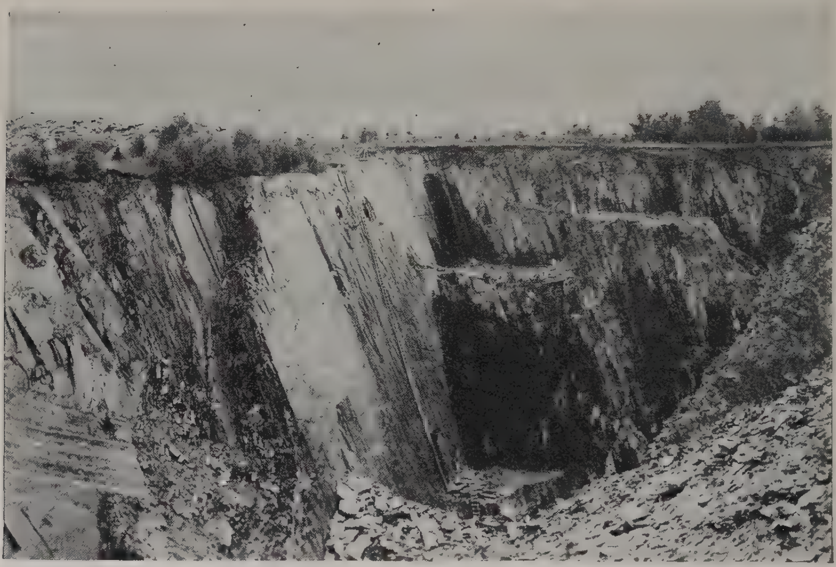


Fig. 248. — Illustrates the occurrence of slates and cleavage. Slate quarries, Brownsville, Me. (U. S. Geol. Surv.)

Hornblende schist. — Hornblende schist is a common and typical variety of schist. It is generally dark green to black, and the parallel prisms of hornblende, if not too large, usually give it a silky luster. Talc schist and chlorite schist are other common schists, in which talc and chlorite are respectively the predominant minerals.

Marble. — Marble is the metamorphic equivalent of the sedimentary carbonate rocks, such as limestone and chalk. Generally the marks of bedding and the fossils are effaced during metamorphism and the material is converted into grains of calcite that are visibly crystalline to the unaided eye. It is, therefore, harder, more compact, with purer colors, and takes a good polish. Just as there are ordinary limestones consisting only of calcium carbonate, and dolomitic limestones containing calcium-magnesium carbonate, $CaMg(CO_3)_2$, in variable quantity in addition to the $CaCO_3$, so we have *calcite* marbles and *dolomite* marbles.

Marble is generally massive and shows no cleavage, even when it has been subjected to great stresses. As rocks go it is relatively plastic and flows under moderate pressures. If for example a dike is inclosed in marble that is forced to flow under differential pressure, the dike, being brittle, will be ruptured and torn apart, and the marble will flow in between the dissevered fragments.

Pure marble is white, and the mottling, banding, and colors shown by ornamental varieties is due to impurities; the red and yellow tones to oxides of iron, the grays and blacks to varying proportions of organic matter. Marble is produced by either dynamic or contact metamorphism.

Serpentine. — This name is given to a mineral, a hydrous silicate of magnesium, and also to a rock largely or entirely composed of it. The rock is usually greenish to black, soft, of a greasy feel, and massive, or without cleavage. Some of the blotched, lighter-green varieties are used as building and ornamental stones. Most serpentines appear to have been made by hydrothermal metamorphism (action of hot waters) on deeply buried masses of igneous rock rich in magnesia, such as peridotite, whereby the magnesium silicates change to this hydrated variety.

CHAPTER XV

MOUNTAINS AND RELATED LAND FORMS

Mountains are of great importance in geology, as they furnish a large part of the information on which the science is based. Dynamic processes, such as stream erosion and glaciation, are especially vigorous and their effects strikingly evident in high ranges. The uplifting of rock masses to great heights has resulted in dissection to unusual depth; and consequently a mountain region affords excellent opportunity not only for descriptive study of rock formations, but also for deciphering the history they record. But if the mountains give aid in solving many problems relating to the Earth, they also present mysteries in themselves. Why have sea floors of remote periods become the lofty highlands of today? What generates the enormous forces that bend, break, and mash the rocks in mountain zones? These questions still await satisfactory answers; but the architectural features of great ranges at least offer hints as to their origin, and are worthy of study for their own sake.

ORIGIN OF MOUNTAINS

Mountains owe their origin to different agencies, and differences in the structure and the plan of mountain units are due largely to this fact. Ultimate causes of mountain building are in large part obscure; but regardless of this fact the principal agencies involved may be stated as differential erosion, volcanic activity, and movements of the crust. Commonly two or more processes combine to produce complex results. However, the discussion will be clarified by a classification that recognizes the dominant agencies.

Residual Mountains

High plateaus suffer differential erosion, and during a late stage in the process some of the more favored residuals may have sufficient height, in relation to their surroundings, to be called mountains. Some of the larger buttes in western United States are examples. The Catskill Mountains in New York represent remnants of an extensive high plateau, the greater part of which has been removed by stream erosion. An early stage in a similar development may be seen in the Grand Canyon of the Colorado, where a number of pyramidal erosion remnants rise from the depths of the chasm (Fig. 249). These are dwarfed by their surroundings; but if they could be placed on a plain they would rise to mountain heights. It is not difficult to imagine that in a later geologic epoch the present youthful plateau will have been dissected thoroughly by the Colorado and its tributaries, and will be represented only by scattered residual mountains separated by plains and valleys.

Plateau blocks from which residual mountains are derived may consist of horizontal rock formations, of homogeneous crystalline rocks, or of folded and faulted strata that were peneplaned by previous erosion. Obviously the resulting mountain forms will be influenced by the original structure. Residual mountains are also known as "mountains of erosion." In a strict classification, however, it must be considered that deep differential erosion is made possible only by great uplift, and therefore movement of the crust is an important factor in producing residual mountains.

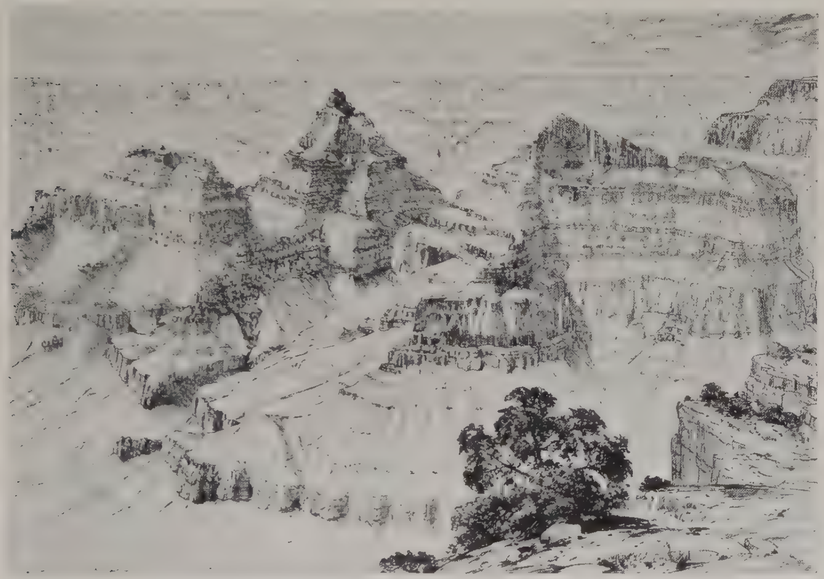


Fig. 249. — Vishnu's Temple, Grand Canyon, Arizona. Mountain-like remnants, left during dissection of Colorado Plateau by Colorado River and tributaries. Part of the level, undissected plateau, many miles distant, appears in background. (U.S. Geol. Surv.)

Volcanic Mountains

Some of the loftiest peaks in the world, such as Chimborazo (20,517 ft.) and Aconcagua (23,393 ft.) in the Andes, and Kilimanjaro (19,321 ft.) in Africa, have been built directly by volcanic action. Many such peaks, however, have their bases on high plateaus, and therefore their true height is much less than the altitudes above sea level indicate. A large number of volcanic islands, such as the various islands in the Hawaiian group, are great volcanic piles, and some of them appear to be seated directly on the deep ocean bottoms.

Volcanic peaks commonly are superimposed on mountains that resulted from other causes. All mountain masses formed directly by igneous extrusion are sometimes called " mountains of accumulation." In areas of widespread volcanism the volcanic materials build up extensive plateaus, such as the Absaroka Plateau east of Yellowstone Park, and the great Columbia Plateau of Washington and Oregon. From units of this kind residual mountains are produced in time by differential erosion.

Mountains Whose Structure Reflects Crustal Movements

The mountain types considered above are important geologically, as they form groups of considerable size in the world today and have had wide distribution during past geologic periods. Differential erosion is one of the major factors in producing mountain relief. However, this factor could not operate unless large areas of the continents were raised high above sea level. Furthermore in most of the dominating mountain units the relief has been conditioned, either directly or indirectly, by localized movements which have caused more or less severe disturbance of the rocks. In some of the young mountain systems much of the local relief was caused directly by these movements; in older belts, which have experienced deep denudation and perhaps more than one rejuvenation by regional upwarping, the original disturbance of the rocks may be important chiefly in guiding erosion. But even if the present mountain relief is due chiefly to differential erosion, any characteristic structure produced by crustal movement has large importance in classifying mountain units. According to the nature of the movements, as indicated in the resulting structure and form, mountains of this general type may be divided into four classes. (1) Dislocation or faulting on a large scale results in relative uplift of rock masses, with or without tilting. Ranges whose structure is produced chiefly by this process are *fault mountains.* (2) Some vertical movements result in arching of the rocks into a general domal form, either nearly circular or somewhat elliptical in plan. *Dome mountains* result from this process. (3) More commonly the forces deforming the crust produce large plications, or parallel anticlines and synclines, giving rise to the structure of *fold mountains.* (4) In most of the great mountain belts we see the combined effects of two or more types of movement, particularly folding and faulting, with complications produced by igneous intrusion. The resulting mountains are of the *complex* type, although locally they may be classified according to the process that has played a dominant rôle.

Fault Mountains

Assume that a system of intersecting fractures, reaching to great depth, divides part of the Earth's crust into blocks or masses of very large dimensions. Mountains may be produced directly by movements of several kinds. (*a*) If the region is initially a high plateau, some of the blocks may be depressed several thousand feet, leaving other blocks in their original positions, to form mountain ranges. (*b*) Regardless of original altitudes, some of the blocks may be elevated to mountain heights, by forces acting largely in the vertical direction, leaving adjacent blocks relatively depressed. (*c*) All of the blocks may move downward or upward, but differentially, so that in the end some stand much higher than others. (*d*) Each of several blocks may be tilted or rotated, one edge being elevated and the opposite edge depressed. The

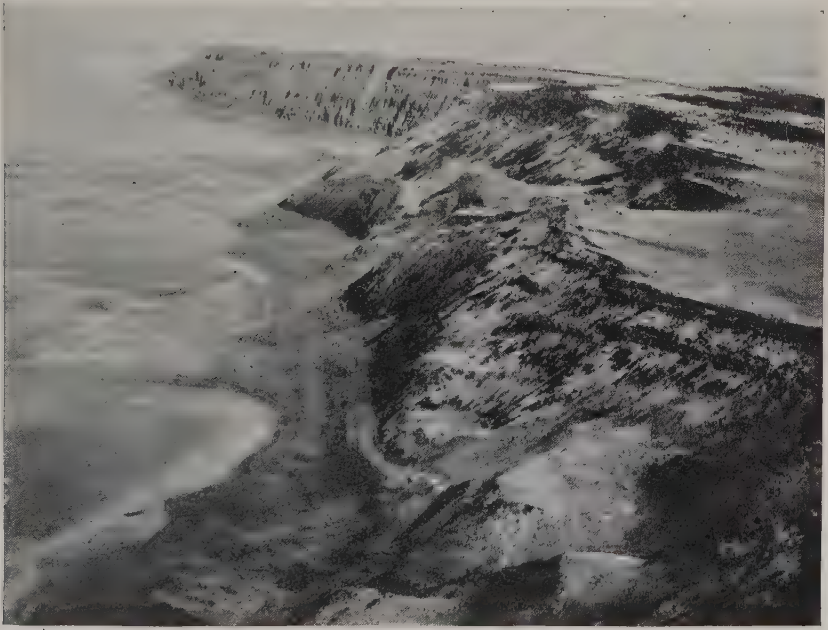


Fig. 250. — An example of fault mountains. A great fault zone, somewhat irregular, marks the base of the range. Downthrown block, at left, covered with lake deposits. Scarp 2000 feet high. Bluejoint Rim, Oregon. (Airplane view by R. E. Fuller.)

faults on which movement occurs may be either normal or reverse. Regardless of the nature of movement, a series of neighboring ranges obviously due to faulting may be called either *fault mountains* or *block mountains* (Fig. 250).

Fault blocks as we actually see them have been more or less modified by erosion. Débris worn from the high masses tends to bury those at low elevation. In time this combination of erosion and deposition may nearly or quite obliterate the mountain relief, especially in a region of interior drainage where all the débris is retained. At a later date, as a result of broad regional uplift perhaps accompanied by change of climate, the original pattern of ranges and intermontane troughs may be etched out by vigorous erosion. Obviously these resurrected ranges are the direct result of differential erosion, and strictly they are residual mountains. However, it is recognized that the original faulting was a strong factor in guiding denudation and continues to be reflected in the mountain forms. Accordingly the original structure is given large emphasis, and it is common to use the designation *fault mountains* or *block mountains* even for units that have been greatly modified by erosion.

The Sierra Nevada of California is a tilted crust block 400 miles in length and approximately 100 miles wide. Its eastern edge has been uplifted two miles or more, to form an abrupt eastward-facing scarp. Roads from the east ascend this precipitous front with difficulty; but west of the crest they descend on a long, gentle slope — the tilted upper surface of the block. In the Great Valley of California, sediments thousands of feet deep have accumulated on the depressed portion of the rotated mass. A great series of fault mountains lies in the Great Basin of Nevada and neighboring states. Ranges in that region are so similar in character that they are known collectively as the Basin and Range System. Some of the ranges are still growing; for within historic time movements have occurred on several of the faults, giving rise to violent earthquakes. As a large part of the Basin and Range region has no drainage to the sea, the mountains are partly buried by accumulations of their own débris.

Dome Mountains

Mountains whose structure reflects crustal uplift of distinctly domal character may be classed together, regardless of size or the exact cause of the uplift. The simplest and best understood are laccolithic domes, made by the bowing up of strata above thick, lens-shaped intrusions of liquid rock. Ordinarily a dome of this kind that is high enough to be called a mountain has lost more or less of its original cover through erosion; and not uncommonly the resistant igneous mass, almost completely denuded, stands within circular or elliptical ridges formed by the upturned edges of the more resistant strata. An excellent example is Bear Butte, one of many laccolithic mountains in the vicinity of the Black Hills (Fig. 180). The Henry Mountains of Utah, classic examples of the type, are a large group of laccoliths in various stages of denudation. But not all mountains of this kind have ideally simple structure. The intruding magma not uncommonly ruptures the covering strata and lifts them irregularly, as in the Moccasin Mountains of Montana.

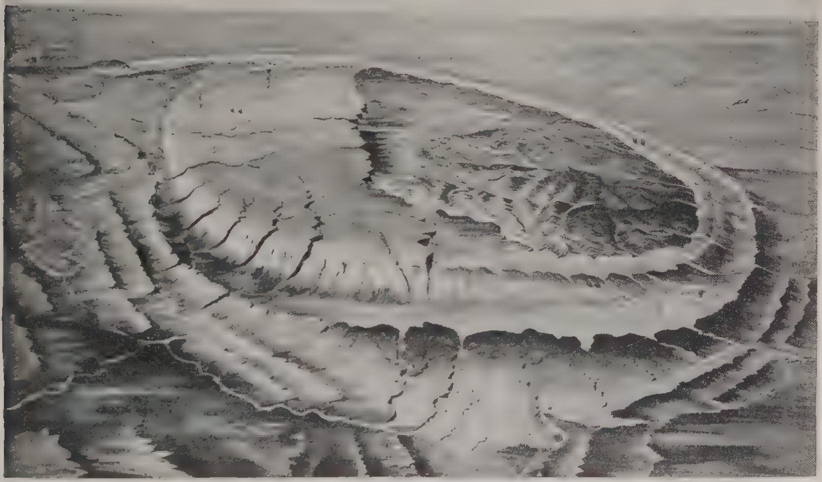


Fig. 251. — Diagram of the Black Hills uplift, South Dakota. View looking north, along the longer axis of the elliptical dome. The wide valley encircling the dome is commonly known as "The Race Track." (Newton, U. S. Geog. and Geol. Survey.)

Dome mountains on a larger scale are illustrated by the Black Hills of South Dakota. In a casual journey through these mountains the traveler may gain an impression of disordered arrangment; but a good map or a block diagram of the entire unit reveals a beautiful symmetry of structure, involving an elliptical area about 100 miles in length by 50 in width (Fig. 251). In the middle and eastern portions the uplifted sedimentary formations have been stripped away, and there the crystalline basement rocks have been carved into ridges and peaks, including Harney Peak (7242 feet above sea level), the highest point east of the Rockies. Around the flanks of the uplift the upturned edges of sedimentary formations, yielding to erosion at different rates, make alternating high and low belts which encircle the uplands. Some of the heavy limestone members still cover the western half of the uplift, forming the "limestone plateau." A careful profile and section, in which the eroded formations are restored to their original positions, shows that the dome is steeper on the eastern than on the western flank, and that the greatest erosion has occurred in the area of maximum uplift. The restored section indicates that the top of the dome was elevated 9000 feet, although the present height above the surrounding plains is less than 4000 feet.

In contrast with a laccolith, the crystalline rocks exposed at the core of the Black Hills were not intruded as magmas, but are much older than the domed sedimentary formations. Therefore the force by which the uplift came about was applied much deeper than any exposed part of the crust. If the cause of uplift was the rise of igneous material, the molten mass was very large, quite symmetrical in form, and very deep-seated. The connection of some igneous activity with the movement is indicated by numerous laccoliths, which form small satellitic domes in the northern part of the Hills. Whatever may have been the part performed by igneous magmas in causing the main uplift, it is very probable that the horizontal pressure by which the Rocky Mountains were folded was an important factor also in shaping the Black Hills dome.

It is evident, on reflection, that the Black Hills uplift did not become an actual mountain group without the work of erosion. If the youngest strata involved in the movement still extended unbroken across the summit, forming a broad, smooth dome, the area would be a small plateau rather than a mountain. It is essential, therefore, to keep in mind the limitations of any scheme of classifying mountains. The limit of size separating dome mountains from plateaus must be somewhat arbitrary; but as the Black Hills uplift is a definite unit of moderate size, with strongly defined boundaries, it seems proper to emphasize the structural form, provided the various steps in fashioning the highlands are kept clearly in mind.

If the Black Hills should be worn down to a peneplain by prolonged denudation, and a later warping movement should reëlevate the Great Plains region several thousand feet, subsequent erosion would be guided by the old structure and a group of residual mountains similar to the present Hills would be produced eventually.

Fold and Complex Mountains

Mountains in which the rocks are strongly folded and broken are commonly described according to their internal structure, regardless of the later chapters in their history. Some old mountain units may be strictly remnants of erosion, and therefore residual mountains. However, if they give evidence that a certain type of crustal deformation attended their early development, these structural characteristics are used as the basis of their classification. For example, the Appalachians have had a long and varied career. The original chain suffered erosion through long ages, and almost or quite disappeared; and the present Appalachian ridges have been etched out after a later upwarping of the area containing the old mountain "roots." Nevertheless, remnants of the original structure are clearly visible, and are recognized as an important feature of the mountain zone. It is well to keep this example in mind throughout the discussion that follows. Deformed rocks are characteristic of all great mountain zones, and the development of this deformed structure will be given prominent attention in outlining mountain history; but it is not to be taken for granted that the deformation gave rise to the present mountain elevations. Mountain *structure* and mountain *elevation* may not have any direct relation to each other. Nevertheless the structure continues as a dominant factor in determining relief because it guides differential erosion.

All the great mountain chains of the Earth include folded sedimentary rocks as a conspicuous part of their structure. These chains, therefore, are sometimes classed together as *fold* mountains, although faulting, igneous intrusion, and other important processes besides simple folding have played some rôle in their origin. Some parts of the Appalachian Mountain belt exhibit fairly regular plication of rock formations, but other parts are complicated by thrust faulting; and as there are all stages of gradation between the simpler sections of this chain and the almost incredible complexity of the Alps, it is clear that *fold* and *complex* mountains cannot be separated as sharply contrasted structural types.

General Characteristics of Fold and Complex Mountains. — From examination of a globe or a world map it is apparent that the prominent mountain belts are elongated generally parallel to the continental margins. This relation is especially striking in the American cordilleras, the Appalachians, the Scandinavian chain, and the great Eurasiatic mountain zone. Each major belt is composed of numerous ranges disposed somewhat irregularly but with the same general orientation. Some of the ranges are nearly straight in plan; but many are strongly curved into the form of great bows or arcs. The Alps, Carpathians, and Himalayas are striking examples of this arcuate type.

Generally the exposed portion of each range is made up in part or wholly of distorted sedimentary formations. Commonly these strata, now on the flanks or even on the highest summits of the ranges, represent deposits in former seas or on deltas and in marshes bordering the sea. Owing to the strong folding and faulting of these strata, followed by planation and dissection through erosion, the full thickness of the sedimentary cover is exposed in many places. In some mountain belts these thicknesses are astonishing; 4, 5, or even 6 miles are by no means exceptional values, and in some mountain areas the total sedimentary sections exceed 40,000 feet. It will occur to some readers that similar thicknesses may be common also outside of mountain zones, but are not known because conditions favorable for their revelation do not exist. However, natural exposures and well records indicate clearly that sedimentary formations grow conspicuously thinner away from a folded mountain belt. Thus the strata in the Appalachians average 20,000 feet or more in thickness along the central axis of the folded tract; but at no great distance to the west the thickness is less than 10,000 feet, and in the Mississippi Valley it is only 4000 to 5000 feet. On the east side of the Appalachians the sedimentary strata do not exist, and it will be shown presently that the deposits never extended far eastward from their present limit. Therefore the excessively thick sediments occupy a long and relatively narrow belt that corresponds closely to the axis of the folded chain. This general relationship exists also in the Rockies, the Andes, the Mediterranean ranges, and other great mountain systems. It is a natural conclusion that the accumulation of abnormally thick sediments had a significant connection with the development of each of these mountain units.

The typical history of ranges with fold structure falls into three general divisions: the *preliminary stages* in which certain processes prepare the place and some of the material for the future mountains; the *period of crustal movements,* in which the folds and related structural features are produced and the initial uplift occurs; and *subsequent stages*, during which the mountains experience various modifications through erosion and repeated vertical movements. This subdivision of the history is useful, if it is understood that the periods are not sharply separated, and that some of the processes involved operate simultaneously. Thus some incidents of folding and thrusting occur long before the accumulation of sediments is complete; and inevitably the modifying influence of erosion dates from an early stage of uplift, while the orogenic or mountain-making processes are active. The growth and decline of every range is the result of a slow, complicated interplay of forces that act either continuously or recurrently during a very long geologic interval. Recognition of general stages in the history serves to clarify a brief discussion; but these stages overlap and merge irregularly into one another.

Preliminary Stages: Development of Geosynclines. — Sedimentary strata in the great ranges consist of conglomerates and sandstones mingled with shales and limestones. It is clear that thick deposits containing coarse sediments must have been laid down near the margin of a land that suffered prolonged erosion. The great thickness of deposits may suggest that sedimentation began in an excessively deep basin. However, there is unquestionable evidence that nearly all sediments involved in mountain folds were laid down in shallow water or at only moderate depths. Accumulation of such deposits to a total thickness of several miles indicates that slow subsidence was continuous or recurrent while deposition was in progress. Moreover, as enormous volumes of coarse sediments were delivered into the subsiding basin repeatedly, the wasting land must have risen continuously or recurrently adjacent to the area of sedimentation. In any case, the preliminary structure that determines the location of a future range appears to be a sinking trough into which the waste from nearby land accumulates to unusual depth. An elongated subsiding tract of this nature is known as a *geosyncline.*

In the Appalachians various features of the strata indicate that conditions within the old geosynclinal trough fluctuated repeatedly. Sandstones and shales with abundant ripple marks and mud cracks are interbedded with thick limestones that contain marine fossils. Such relations imply a shifting coast and considerable variation in depth of water. In fact at some periods the sea gave place to great delta plains or to enormous swamps in which materials for coal beds accumulated.

From study of the sedimentary sections in the Appalachians it is clear that the coarser sediments are on the east, and to the west these give way to marine shales and limestones. Therefore the land from which sediments were eroded lay to the east of the geosyncline. A narrow belt of ancient rocks near the present coast presumably represents the western edge of the former land; but it must have extended far eastward, over the area of the present continental shelf, in order to have volume adequate to explain the vast quantities of sediments it supplied. The name " Appalachia " is applied to this ancient land of unknown extent (Fig. 252). The sea that lay west of it, covering much of the present Mississippi Valley region, was shallow and fluctuating.

Similar histories have preceded the making of other great ranges. The Rocky Mountains grew up from a great geosyncline that stretched from the Gulf of Mexico to the Arctic, with highlands to the west. The old land that furnished sediments for the Alpine geosyncline lay to the north of the Alps. Strata folded in the Caucasus were derived from lands to the south while seas stretched northward over Russia. Thus it may be accepted as a general principle that on one side of the mountain zone lies an area of much older rocks, the source of the folded sedimentary deposits. The time occupied in the accumulation of sediments in the geosyncline extends over long geologic periods.

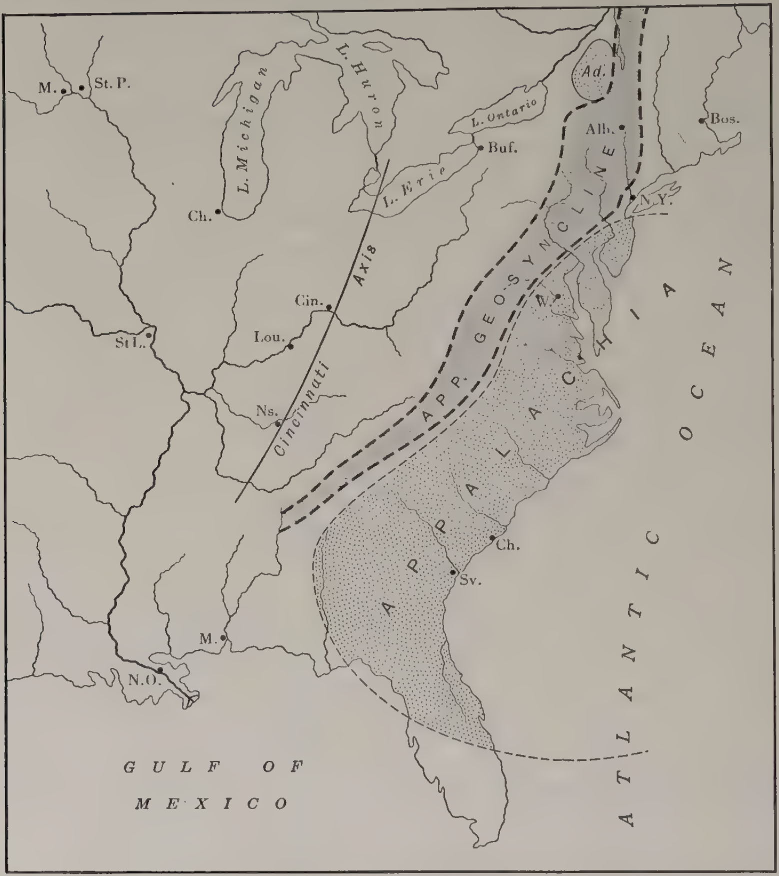

M.
St. P.
L. Michigan
L. Huron
L. Ontario
L. Erie
Ad.
Alb.
Bos.
Buf.
Ch.
N.Y.
Cincinnati Axis
APPALACHIAN GEOSYNCLINE
Cin.
Lou.
St L.
Ns.
W.
APPALACHIA
Ch.
Sv.
M.
N.O.
ATLANTIC OCEAN
GULF OF MEXICO

Fig. 252. — Map showing the situation of the Appalachian geosyncline and of the old land Appalachia. *Ad*, mass of the Adirondacks.

Period of Crustal Movements. — The period of relatively quiet preparation, of long-continued erosion and sedimentation and slow movements of land surface and sea bottom, gives way to a period of greater activity in which the Earth's outer shell yields to powerful lateral pressure. By this pressure the accumulated load of sediments is thrown into folds, crushed and mashed together into the disordered arrangement characteristic of the great chains. The process and its results thus simply stated are in reality very complicated, with different phases and with divergent features in different regions.

It is clearly evident, from the structural features found in complex mountains, that crushing of the old geosyncline and its burden of sediments was performed by forces acting in a lateral direction, tangential to the Earth's surface. Thus in zones of most intensive folding, the folds not only become closed so that their limbs are in contact, but they are even more severely compressed, with mashing of the beds and the production of very complicated structures. This is illustrated by sections in the Alps and in the Appalachians (Fig. 253). Considering the scale of these sections it is impossible to imagine the formation of such structures except by transverse compression of great magnitude.

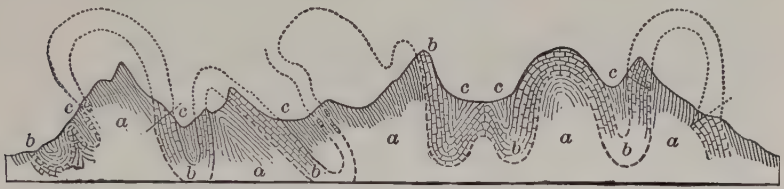


Fig. 253. — Section across the Santis Alps, N. E. Switzerland (after Heim, somewhat modified). *a*, shales, breccias; *b*, massive limestone; *c*, shales, thin limestones.

The varied phenomena of folding shown in the mountains may be imitated by lateral compression applied to a sequence of artificial strata composed of some plastic substance, such as wax or clay, placed upon one another. If some of the layers are made somewhat harder and stronger than others, and the series is laid in a firm trough or box, one end of which may be forced inward by turning a screw, the stronger layers tend to form large and simple folds, whereas the weaker members are distorted in a more complicated way. With continued pressure the folds may be overturned and broken. The displacements and dislocations, the folding and faulting of the strata, produced in miniature by this method, are similar to those observed on a great scale in the mountain ranges (Fig. 254).

Cleavage. — It has been shown that cleavage in metamorphic rocks, such as slates, is produced by great pressure, and that the planes of cleavage are developed at right angles to the direction of pressure. Many of the rocks of the great ranges in the zone of intensive folding have been turned into gneisses, schists, or slates, depending on their original composition and other factors. This alteration becomes more evident as the inner portions of the compressed masses are exposed by erosion. Observation shows that the planes of cleavage usually stand at high angles and not uncommonly are vertical, whereas the strike of the cleavage planes is generally parallel to the axis of the range. The direction of the compressive force, thus indicated by the cleavage, is the same as that shown by the folding.

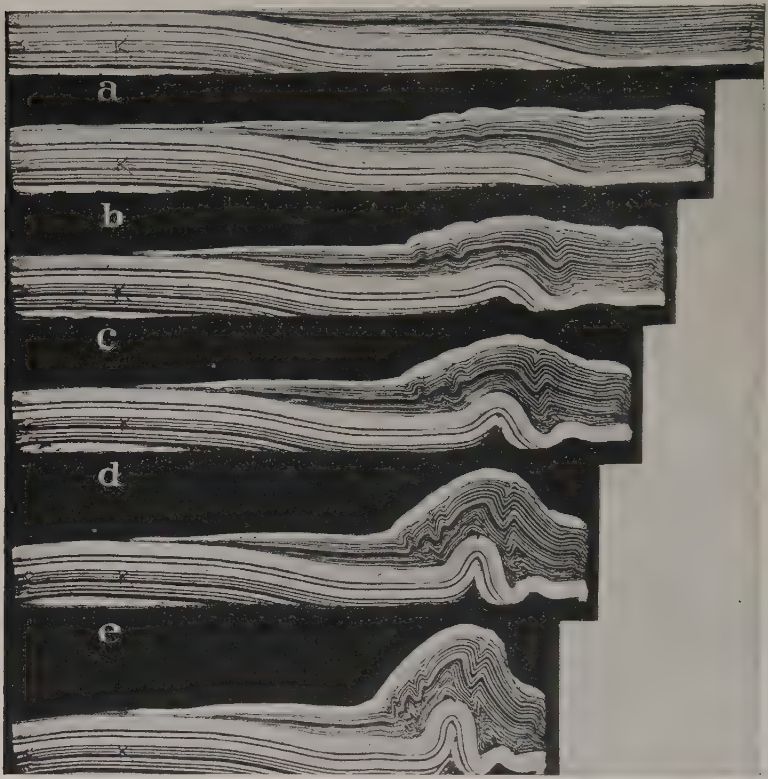


Fig. 254. — Layers of wax and plaster folded by lateral pressure, imitating structures found in mountain ranges. The thrust is from the right, and in successive layers from *a* to *e* the amount of shortening can be seen. (Willis, U. S. Geol. Surv.)

Amount of Compression. — The magnitude of the forces involved and of the masses operated upon is indicated by the amount of compression which investigation shows has actually occurred in some of the great folded belts. In the Appalachians, estimates of 40 to 50 miles, and in some sections even more, are given for the distances the original width of the strata in the geosyncline has been decreased by the mashing together of the mass. In other words, if the folded strata in Pennsylvania, which resemble a crumpled blanket, could be smoothed out toward the southeast, their extent would be increased sufficiently to cover the state of New Jersey. The Rocky Mountain structures represent a comparable amount of shortening. Thus the original breadth of the geosynclines has been diminished by tens of miles, and in the zones of intensive folding and mashing the reduction has been one half or more.

Variations of Folded Structure. — It is to be expected that the results of folding should differ considerably in character between separate ranges, or between distant parts of the same range. Differences in thickness of sediments, in proportions between strong and weak formations, in the form of old rigid masses that transmitted the thrust, and in severity of the lateral forces, are reflected in the individuality of mountain folds. The Jura Mountains, a small member of the Alpine system, furnish classic examples of symmetrical, upright folds (Fig. 255). These folds were produced far out in front of the Alps proper, in a relatively thin sheet of strata, as an incidental effect of the forces that deformed the greater Alpine zone. The Appalachians present a wider variety of fold structures. In the slate and anthracite regions of eastern Pennsylvania the folds are closely compressed, many of them to the isoclinal stage, and the axial planes are strongly overturned toward the northwest. Farther west in the state the folds tend to be open and upright; and the deformation dies out gradually westward. Going to the south, through Virginia and Tennessee and into Alabama, we find that many of the folds were ruptured by the severe compression and developed into thrust faults (Fig. 256). This kind of complexity is especially pronounced in the Alps.

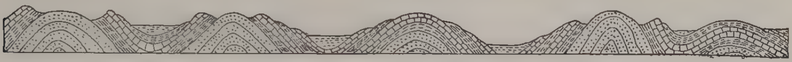


Fig. 255. — Diagrammatic section across part of the Jura Range, showing simple structure and symmetrical folding.

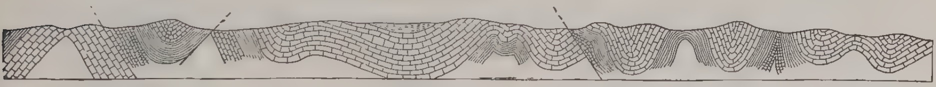


Fig. 256. — Section 12 miles long illustrating Appalachian structure near Greeneville, Tenn. (Slightly modified from Keith and Willis.)

Alpine structure is characterized by great folds that have been pushed over to a horizontal attitude, and by flat thrusts that are related to these overturned folds. These features are developed on an unprecedented scale, with the result that the Alps consist of a series of great rock sheets, driven one over another and overlapping like the shingles on a roof. Erosion has cut valleys and " windows " through the sheets, exposing the entire series; and in parts of the Alpine area nearly the whole of one or more sheets has been swept away, leaving remnants of old rocks to form isolated peaks standing on younger rocks that were overridden and covered during the thrusting movement. Isolated peaks that have this anomalous relation are called "mountains without roots." The Matterhorn and the Mythen are famous examples.

Mountain Elevation. — There is a natural tendency to assume that mountains rise to greater heights continuously with folding and thrusting, as a logical result of crowding excess material into a narrow zone. For a long time, indeed, this conclusion was taken for granted, and attempts were made to estimate the original height of eroded ranges by restoring the folds from study of the eroded limbs. As the steps in mountain history become clearer, however, it is found that much of the actual elevation occurred at a distinctly later time than the folding and thrusting. After the Rocky Mountain deformation in early Tertiary time, the folded and faulted area was eroded to a nearly even surface at a low altitude; and the present great heights in the Rockies are due to vertical uplift in late Tertiary time. Similarly, after much of the thrusting and folding was complete the Alps had only moderate height, and the sea washed the flanks of the range both on the north and on the south. In very recent geologic time a movement of vertical uplift carried the Alpine summits to great height. The Andes and the Himalayas have had a similar history.

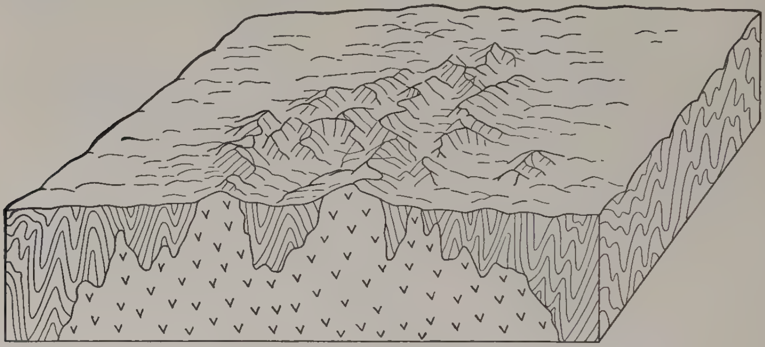


Fig. 257. — Diagram to represent the granitic core of a mountain range after prolonged erosion. The granite batholith cuts across folded and metamorphosed sediments.

Rôle of Igneous Agencies. — Although the making of mountain structures by compression appears to be independent of direct igneous action, and some ranges contain little or no visible igneous rock, nevertheless an upwelling of magma to produce both intrusive and extrusive bodies has been a common incident in mountain making. The effect of this action is to modify the structure due to folding and faulting, and by the addition of large massive bodies to increase the rigidity and strength of the mountain zone. Probably the most effective way in which this happens is by the intrusion of great batholiths, usually composed of granite, into the inner, lower portion of the range. A granite intrusion of this nature may become exposed later by deep erosion, and is then spoken of as the " granite core " of the range (Fig. 257). Intrusion of the heated magma, combined with the folding and mashing of the strata, causes profound metamorphic effects over wide areas. As great intrusive bodies of this kind generally cut across the folds and thrusts, it appears that the intrusion occurred late in the period of orogeny and had no direct connection with the compressive forces. Invading granitic masses of this character are conspicuous features of the Coast Range in western Canada and of the Sierra Nevada.

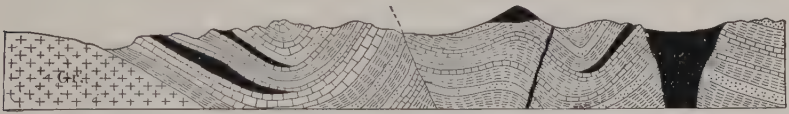


Fig. 258. — Section illustrating intrusions of igneous rock (black) in a folded and dislocated mountain region. *Gr*, edge of a granite batholith.

Intrusions of molten magmas not only make great batholiths, but pressing upward along belts of weakness caused by folding and faulting, they form intrusive sheets, laccoliths, dikes, and other bodies (Fig. 258); or, attaining the surface, they are extruded as lava flows or give rise to violent volcanic action. During the formation of the Rocky Mountains, Wyoming and Montana were the scene of great volcanic activity, the dying phases of which are still evident in Yellowstone Park.

Subsequent Stages. — Although mountains may be classified according to the crustal movements involved in their formation, it is not to these processes alone that the mountains in their present form are due. Hand in hand with uplift of the masses goes the work of erosion, that mighty chisel of Nature, which shapes and carves them into the forms familiar to us. In a sense, therefore, all mountains are *residual*, as erosion starts with the beginning of uplift, proceeds during the whole period of orogenic activity, and continues so long as the highlands exist. During the subsequent stages the history of a range consists chiefly of progressive changes due to erosion, by which the mountains reach maturity and old age. This development is discussed on later pages.

Geologic Date of Mountain Making

The geologic period in which a geosyncline has been crushed, and the strata compressed into mountain folds, is fixed by determining the age of the latest strata involved in the folding, and of the oldest undeformed beds that lie upon the disturbed rocks about the mountain flanks. Obviously the folding is younger than any of the folded or disturbed beds and older than any that are undisturbed (Fig. 259). The closeness of dating by this method evidently depends upon the length of the interval between the dates of deposition of the two sets of strata. Thus if the youngest folded rocks are of late Triassic age, and the oldest undeformed sediments were deposited early in the Tertiary, the folding may have occurred either in the Jurassic or in the Cretaceous period. The Appalachian folding is dated rather closely, as Permian rocks are affected and Triassic rocks are not.

Mountain making by any process is dated according to the same general principles. The latest episode of uplift in fault mountains cannot be older than the youngest rocks affected by the faulting.

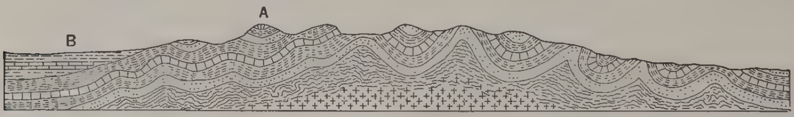


Fig. 259. — Illustrating determination of the date at which deformation in a mountain zone occurs. *A* are the youngest surviving strata that were involved in the folding; the folding is younger than these. *B* are not concerned in the process, and hence the folding is older than they are.

The Ultimate Cause of Mountain Making

When it was believed that the Earth consisted of a relatively thin crust resting on a highly-heated liquid interior, it seemed easy to explain the origin of mountains by assuming that there was a regular contraction of the Earth's mass from loss of heat, and, as this contraction was greater in the heated interior than in the cold outer crust, the latter was folded up as it gradually sank upon the shrinking core, very much as the skin wrinkles upon a drying and contracting apple. This view can no longer be held, because the Earth is known to behave as a solid, rigid body; the interior cannot be wholly, or even chiefly fluid, in the ordinary meaning of the word.

Nevertheless a view commonly held to account for crustal deformation assumes contraction by loss of heat. This is a survival of the idea mentioned above, but changed to accord more nearly with later knowledge. It assumes that the Earth is solid and rigid but very hot within, and that progressive loss of heat causes slow shrinkage below a comparatively shallow depth. An English scientist has shown that this mechanism would account for a large amount of lateral thrusting in the outer shell throughout geologic history, provided we make certain reasonable assumptions as to the temperatures in the interior. Another form of the contraction hypothesis discards the idea of cooling, but assumes that the enormous pressures deep in the Earth cause matter to increase in density and decrease in volume. The net effect would be the same as if cooling occurred, as the outer shell, unchanged in volume, would collapse on the shrinking core, with consequent folding and thrusting. In either form the hypothesis encounters grave mechanical difficulties; among them, the necessity for explaining localization of deformation, during any one period, in only two or three widely separated geosynclines. It seems more probable that uniform contraction of the Earth would cause moderate failure of the crust in many zones, instead of violent deformation in a few.

Within the last few years some geologists have suggested that whole continents may shift horizontally through long distances. It is claimed, for example, that Africa moved northward against the old Mediterranean geosyncline and crushed it to form the Alps and neighboring mountains; and that the great folded chains of Asia were caused by southward shifting of that continent. It is urged that no other explanation will suffice in view of the stupendous shortening recorded by mountain folds and thrusts. But even if we should admit the moving of continents, the fundamental problem of crustal deformation would remain unsolved so long as the ultimate forces and conditions to cause such movement are wholly unknown.

It must be admitted, therefore, that the cause of compressive deformation in the Earth's crust is one of the great mysteries of science and can be discussed only in a speculative way. The lack of definite knowledge on the subject is emphasized by the great diversity and contradictory character of attempted explanations. It is a fascinating problem, but lengthy discussion of its various aspects has no place in this volume. The facts and relationships of mountain structure present a large field of study in themselves, aside from the problem of ultimate forces.

Fault mountains that are due chiefly to thrust faulting present problems similar to those encountered in mountains with folded structure, as the fundamental cause is enormous lateral compression. Some steep faults that bound mountain blocks appear to be related to irregular shifting of magmas at depth, during the formation of intrusive or extrusive igneous bodies. Still others are formed in or near belts of severe folding, by irregular local twisting of the crust during the period of compression; and as fault mountains made in this way are merely incidental effects of the larger deformation, they become a part of the greater problem. Many dome mountains certainly owe their formation to igneous intrusion, and therefore the problem of their ultimate cause is closely linked with the problems of igneous activity.

EROSION IN MOUNTAIN REGIONS

Earlier Stages of Erosion. — So long as the mountain-making forces are at work, a mountain range grows in so far as its structure is concerned. Whether it actually rises in height or not depends on the adjustment between (1) vertical uplift which tends to make it rise, and (2) the work of erosion which tends to cut it down. Always during the formative period this struggle goes on, and the height of the range at any time is a function of these two forces. When the movements cease, then denudation has full sway and, ultimately, with the lapse of time and provided no renewal takes place, the range must be cut down, baseleveled, and extinguished by the relentless agents of erosion. In this process various stages are to be distinguished. When the range is at its maximum elevation the erosive agencies are most severe; to the work of running water on steep slopes is added very commonly the effect of frost, snow, and ice. The mountain masses are profoundly graved and acquire rugged peaks and towering rock pinnacles, alternating with deeply scored valleys.

Later Stages of Erosion. — If erosive processes continue their work of degrading a mountain mass, unhampered by further uplifts, the range gradually passes into a mature stage. The sharp peaks and asperities tend to disappear, the valleys to widen. The progress of the work goes more slowly as the slopes lessen, and as the resistant metamorphic and crystalline rocks of the inner core are reached. Thus a maturely dissected range presents rather smoothly rounded forms and outlines (Fig. 260), which contrast sharply with the angular features of a young range. As they wear down more and more and pass gradually into old age, we find these mountains in humid climates composed of massive rocks, of schists, gneisses, and granites, rather than of the limestones, sandstones, shales, and lavas of ranges in the earlier stages. In arid climates, on the other hand, limestones outlast the granitic rocks because they are far less vulnerable to the mechanical weathering that characterizes dry regions.

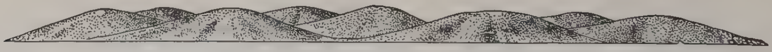


Fig. 260. — Characteristic forms and outlines of late-mature mountains.

Final Stage: Peneplanation. — Ultimately, provided no new upwarping movements occur, the mountains will disappear and the region they occupied will be reduced nearly to baselevel. Since, however, the process of erosion goes on more and more slowly as the slopes lessen, it would evidently require an enormous lapse of time to bring down actually to baselevel a mountainous tract, and we have no proof that this has ever in fact occurred. But we know that some areas have been reduced to low, almost featureless country; in other words to peneplains. Such country may still be diversified by scattered monadnocks projecting above the general level which, on account of their more resistant composition, or possibly because of their position far from large streams, have not been reduced like their neighbors (Fig. 261).

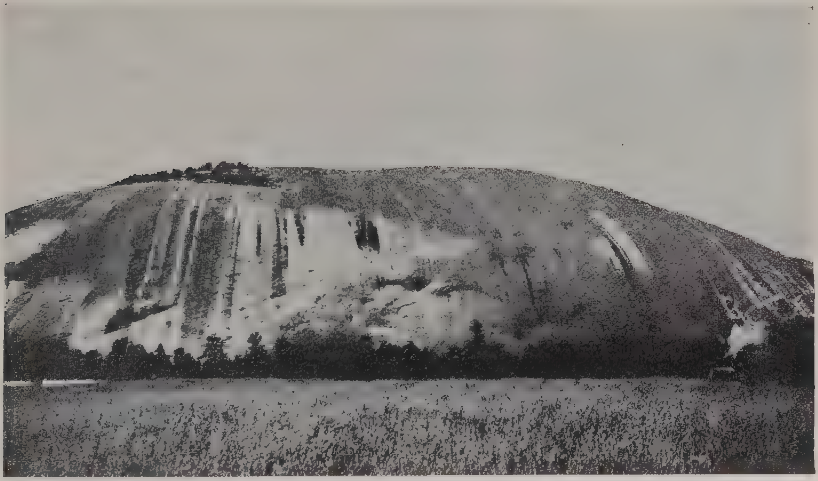


Fig. 261. — Stone Mountain, De Kalb Co., Georgia. A monadnock composed of granite which rises above the surrounding plain of erosion. (Geol. Surv. of Georgia.)

Reëlevation; Complexity of Mountain History. — If mountains were forever extinguished by the peneplanation at the end of a cycle, their history would be comparatively simple. But though the surface of the land may be smoothed out by erosion, the structure below the surface remains. The tilted and folded strata of unequal hardness and the igneous intrusions injected into them are still there, having disappeared as relief features because of peneplanation, but needing only a second uplift of the land to bring them again into prominence. No new deformation of the rock is required. A simple upwarping of the beveled surface, such as has occurred again and again throughout the Earth's history, accomplishes the result. The sluggish streams are rejuvenated;

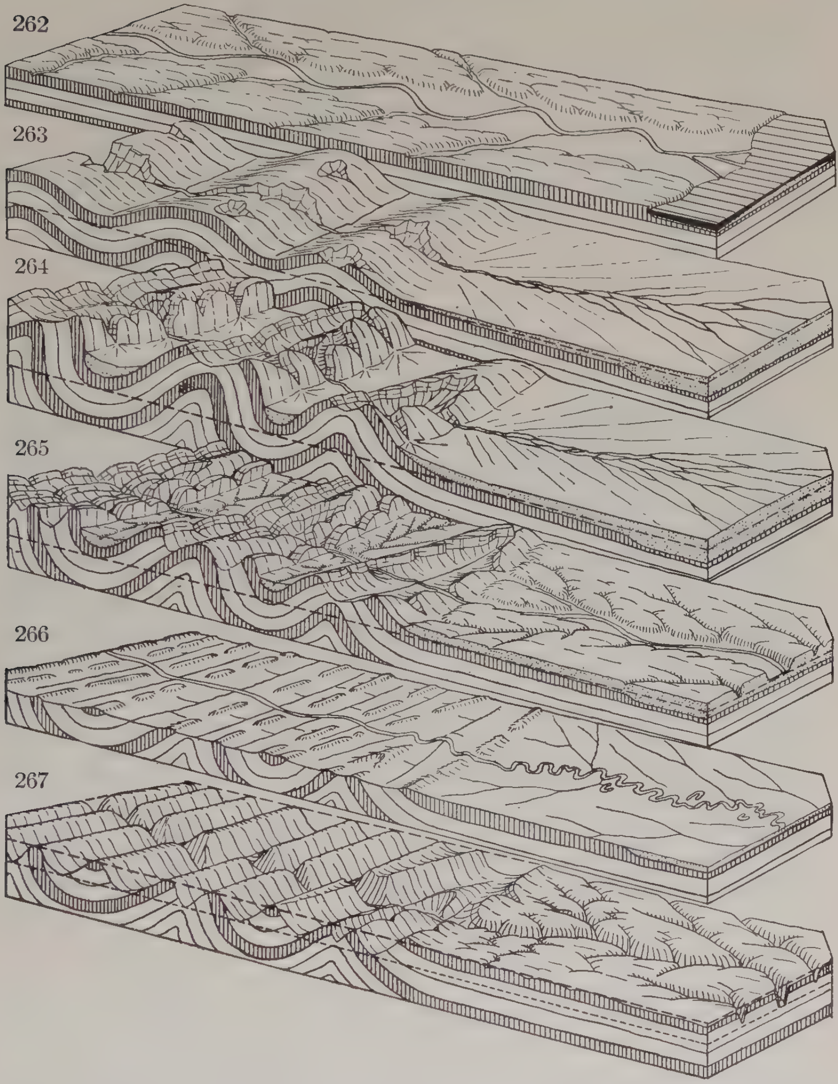
they begin to cut actively, and *they cut the weaker rocks most rapidly,* leaving the resistant masses once more projecting above the surface. The mountains are thus etched again into relief, their new height depending entirely upon the amount and rate of the new uplift.

The steps in mountain history outlined in this general way require a more detailed systematic treatment, since many of the most significant land forms are developed on the disturbed rocks in mountain zones. An important part of stream development was postponed from Chapter IV, because it can not be explained intelligently until the principal facts of mountain history have been outlined. The development of streams in a folded region will now be considered. In connection with this account, constant reference should be made to Figs. 262 to 267.

The Cycle of Stream Erosion in Folded Strata

Initial Stage. — Let the scene be set with a low land developed on a series of horizontal sedimentary rocks, alternately resistant and weak (Fig. 262). The area is drained eastward by one main stream which reaches the sea over a well-developed flood plain and delta. The tributaries, because the rock is horizontal, form a typical dendritic pattern.

Stage of Early Youth; Antecedent Streams and Structural Valleys. — As folding begins, this section of the crust is thrown into three gentle anticlines (Fig. 263), of which the first plunges northward. The increased gradient attendant upon the folding brings the sluggish main stream to life, greatly increases its erosive power, and allows it to trench downward as rapidly as the anticlinal arches are uplifted. A stream which is strong enough thus to maintain its former course in spite of gradual uplifts across its path is said to be *antecedent.* Columbia River is antecedent to the great uplifted arch of the Cascade Range in Oregon and Washington. The tributaries, however, smaller and weaker than the main stream, fare less well; they are blocked off by the folds and are extinguished. But new drainage lines are provided by the synclinal troughs. Concentration of drainage takes place in each of these, and pairs of tributaries are thus developed at right angles to the main stream, as consequents upon the folded surface. In this way the dendritic stream pattern is extinguished and a right-angled, or *trellis* pattern, takes its place. The synclinal tributary valleys are termed *structural valleys* because their position is determined entirely by rock structure. At the same time, rapid run-off down the slopes of the anticlinal limbs forms consequent tributaries of the second order, which rapidly excavate deep gorges. Directly these tributaries cut through the surface bed (which in this case is of resistant rock), and penetrate the weak bed below, they



Figs. 262–267. — Ideal cycle of stream erosion under a humid climate and in folded strata. Each block is oriented so that its right end faces east and its long upper edge faces north.

Fig. 262. — Initial stage, showing a land mass in old age.

Fig. 263. — Stage of early youth, showing the beginning of folding and the development of an antecedent main with consequent tributaries in a trellis pattern.

Fig. 264. — Stage of later youth, showing valley fill, extensive breaching of the anticlines, and the development of subsequent valleys bordered by hogbacks.

Fig. 265. — Stage of maturity, showing dissection of the earlier valley fill, decrease in relief, and the development of rounded ridges within the anticlines.

Fig. 266. — Stage of old age, showing the weak rocks reduced nearly to baselevel, leaving only monadnocks of resistant rock for the stream to cut. Essentially a peneplain.

Fig. 267. — Stage of maturity in a second cycle instituted by regional uplift and rejuvenation. (See page 342.)

begin to cut the latter, undermining the former, and the valleys are quickly widened. This process is characteristic of the early stages of the cycle, in that the weak rocks are attacked as widely as possible, while the hard rocks, escaping direct attack, fall by undermining. The main stream crosses the anticlinal arches through *water gaps* of its own cutting (Fig. 268), and the greatly increased volume of débris in transport is deposited in part upon the uplifted coastal plain. That the stream system is confronted by a long period of cutting is indicated by the volume of rock above the new baselevel (indicated by a dashed line, Fig. 263).

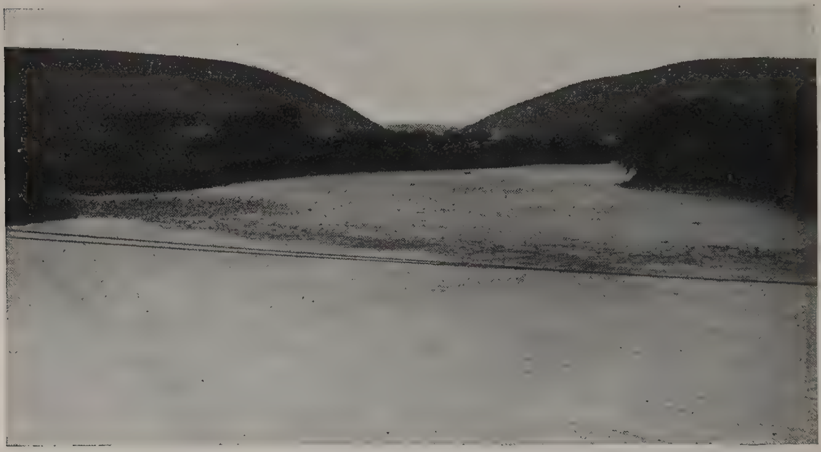


Fig. 268. — Lehigh Water Gap as seen from Slatington, Pa. (Barrell.)

Stage of Later Youth; Hogbacks and Canoe-shaped Valleys. — In this case the folding and general uplift is depicted as having now reached its climax (Fig. 264). The crests have thereby attained their maximum elevation above baselevel (dashed line, Fig. 264). But because much erosion has taken place since the uplift began, it is clear that the actual mountains can never be as high as might be inferred at a later time by projecting upward the stumps of the anticlinal limbs. Steepening of the anticlinal slopes (compare Fig. 264 with Fig. 263) has resulted in such greatly increased erosion by the tributaries of the second order that the streams in the synclinal valleys, whose gradients have not been steepened much since folding began, are not able to carry away the waste contributed to them. The synclinal valleys are therefore partly silted up with waste in the form of coalescent alluvial fans. The ridges formed by the outcropping hard bed resemble steep cuestas. Such ridges of steeply-dipping rocks are termed *hogbacks*. They are common in the Colorado Rockies, the Bighorn Mountains, the Black Hills, and many other anticlinal mountain masses (Fig. 269). Whereas the anticline in the center (Fig. 264) has been breached along its entire crest, the anticline to the east is only partially breached, and that where its axis is highest. Since this is a plunging fold, its axis descends to the north, and hence the tributaries of the second order have lower gradients. This being the case, their cutting power is far less than that of their steeper neighbors, and the resistant bed which caps the anticline is therefore much less readily cut through. This brings about the excavation of a *canoe-shaped valley*, the prow of the canoe pointing down the plunge of the anticlinal axis whereas its sides are formed by the anticlinal limbs. The canoe gradually widens as the retaining escarpments retreat. The whole gaunt structural skeleton is now dissected out in greatest relief, and the stream system has much cutting to accomplish before it can remove all of the rock above baselevel and carry it into the sea.

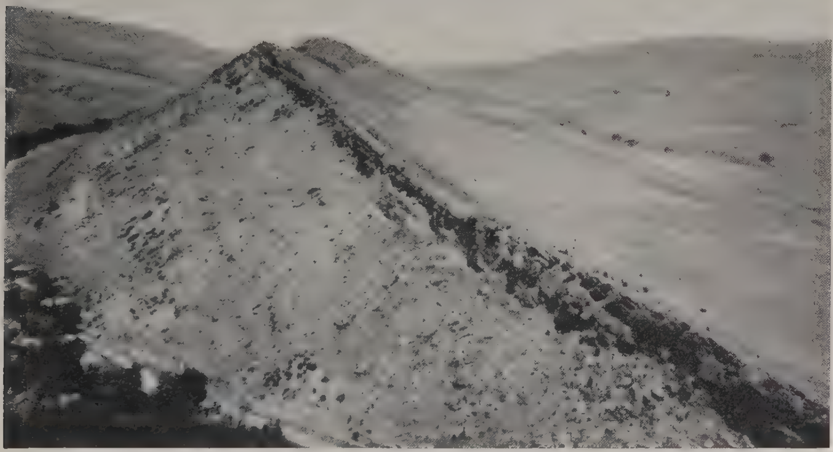


Fig. 269. — A hogback near Morrison, Colo. (U. S. Geol. Surv.)

Stage of Maturity. — The weak strata continue to receive the most vigorous direct attack by the streams, and the resistant strata fall block by block as they are sapped and undercut. Both the canoe-shaped valleys and the others cut in the anticlines have been widened thereby (Fig. 265). Erosion of the soft strata in the middle anticline has now completely laid bare the rounded back of the lower resistant bed, which has already been breached to form an inner canoe-shaped valley. Inspection of Figs. 264 and 265 shows the appearance and development of other such rounded ridges at the hearts of the anticlines.

The whole land surface by this time has been reduced appreciably toward baselevel (dashed line, Fig. 265; compare Fig. 264). In fact, the heights have been lowered sufficiently to decrease greatly the supply of waste to the tributaries. The latter, freed from a part of their load of sediment in transport, are able again to devote a part of their energy to downcutting, and they begin to dissect their previous deposits in the synclinal valleys and on the coastal plain, cutting them into pronounced terraces. The main stream has now come so close to baselevel that it can cut down but little further. Hence it is easily diverted, and it begins to meander widely, broadening its valley by lateral planation. In this it is slowly followed by the tributaries in the synclines.

Stage of Old Age; Beveling of Resistant and Weak Rocks Alike. — The increasing predominance of lateral planation during late maturity and old age has resulted in the virtual cutting away of the remaining highlands. The reason is simply that all of the soft rock has already been worn down essentially to baselevel, leaving nothing but hard rock for the streams to cut (Fig. 266; note that the dashed line has disappeared because it practically coincides with the old age surface). Because of the great resistance of the only erodable rock and the low gradients, the process of cutting is almost infinitely slow. Figure 266 depicts a peneplain carrying elongate monadnocks formed by outcropping edges of the resistant beds, together with one rounded ridge in the plunging anticline. The structure is essentially beveled, hard and soft rock alike. The cycle is virtually ended; from now on the streams can work only to destroy the monadnocks, with scarcely appreciable results.

Rejuvenation of the Fold Structure. — Suppose now that the peneplain shown in Fig. 266 suffers uniform uplift; in other words, that the former baselevel (upper dashed line, Fig. 267) is abandoned in favor of the new baselevel (lower dashed line). This permits the stream system to renew the work of excavation which it had completed under the old conditions. The weak rocks are again etched away from the resistant rocks, which are left standing as prominent ridges. The tops of these ridges are beveled flat, and they are worn down so slowly that during all of the early part of this new (second) cycle they preserve the level of the former peneplain from which the intervening weak rocks have been cut away. These ridges are said to be *peneplain remnants*, and can be formed only by uplift and the beginning of a second cycle following the completion of an earlier one. In an analogous manner, the highest ridges of the folded Appalachian ranges are remnants of a former peneplain. Their flat beveled tops betray the fact that they once stood at or close to baselevel and that they were later etched into relief following a great uplift.

In Fig. 267 the weak rocks in the two synclines are temporarily preserved as ridges because protected by narrow caps of resistant rock. The latter however must disappear in time by being undermined, and the synclinal ridges will then be quickly destroyed. The completion of the second cycle will see a baseleveled surface (developed in the plane of the lower dashed line) much like that of Fig. 266. There will be two chief differences, however: (1) Monadnocks will be fewer because of the complete removal of the upper resistant layer from the two synclines. (2) The remaining monadnocks will be farther apart because the hard layers will be intersected *farther down their limbs* by the plane of the new baselevel. Thus the hills, strong bastions in appearance, are in reality slowly moving down the dip of the resistant strata of which they are composed.

Adjustment of Streams to Structure. — A close study of Figs. 262–265 shows the remarkable fact that although the initial tributaries develop as consequents in the synclines, they steadily decrease in importance until in old age (Fig. 266) they are almost completely extinguished. The further fact appears that as the resistant caps of the anticlines are breached, subsequent streams (page 73) are developed in the weak rocks below, parallel to the synclinal consequents. As the cycle progresses these subsequents gain in importance, so that before the end they have become the most important tributaries. This growth of the subsequents at the expense of the consequents is called *adjustment to structure* and is explained by the simple fact that the former can cut more rapidly in thin weak rock beds than the latter are able to cut in the resistant rocks through which they flow. From this the universal principle may be set up that *stream systems are constantly forced to adjust their courses so as to flow as much as possible on weak rock and as little as possible on resistant rock.* Drainage developed in folded strata rarely reaches an adjusted condition before late maturity or old age. The more or less complete adjustment of the larger streams in the folded Appalachians is illustrated in Fig. 270. From the trend of the trellised stream pattern the strike of the folds can be readily inferred.

Superimposed Streams. — Certain streams are so completely out of adjustment and pursue courses so wholly regardless of structure that some special set of conditions must have brought about this inharmonious relationship. Suppose the seaward portion of a peneplain developed on folded strata of unequal resistance to be depressed slightly. During the resultant submergence, marine deposits are laid down almost horizontally upon the drowned erosion surface, forming an angular unconformity (page 289) with progressive overlap (page 203). When now the monotonous submarine plain is uplifted, consequent or extended streams develop on the gently sloping surface, exhibiting a dendritic drainage pattern (Fig. 271, *A*). In time the streams cut down through the unconformity into the different material and structure below. When the overlying sediments have been stripped away, the streams are

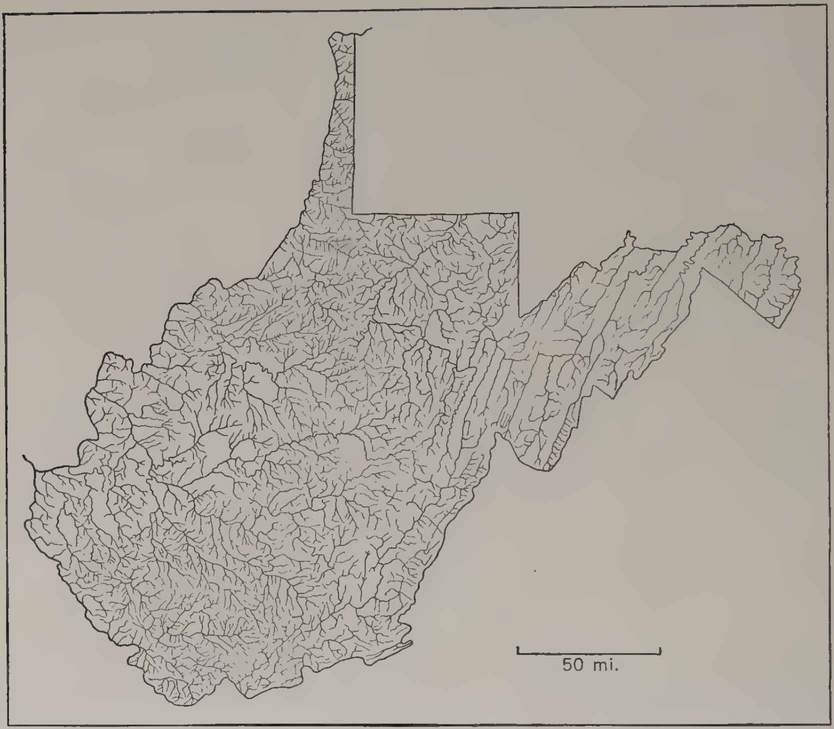


Fig. 270. — Outline map showing stream patterns in West Virginia. In the central and western regions the rocks are nearly horizontal and the pattern is dendritic. In the east the rocks are strongly folded along NESW axes and the resulting pattern is trellised.

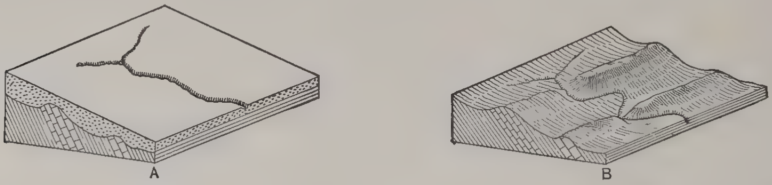


Fig. 271. — Development of a superimposed river. *A*, course determined by initial slope on a layer of sand and gravel; *B*, latter removed by erosion and river pursuing its course without regard to underlying rock structure.

found to be following unadjusted courses *inherited from the overlying unconformable strata* (Fig. 271, *B*). Such streams are said to be *superimposed.* Streams may be thus " let down from above " from glacial, volcanic, eolian, and alluvial deposits as well as from marine deposits. In any case, gradual adjustment will take place as the cycle progresses.

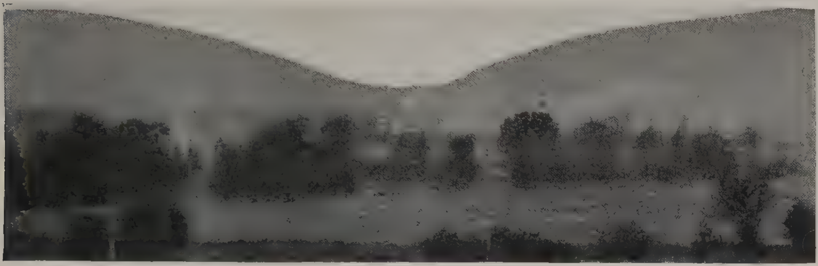


Fig. 272. — Wind Gap in Kittatinny Mountain at Pen Argyl, 11 miles north of Easton, Pa. (Barrell.)

Wind Gaps; Piracy by Subsequent Streams. — A water gap (Fig. 268) abandoned by its through-flowing stream is called a *wind gap* (Fig. 272). Most of the wind gaps which have been studied are the result of *piracy* by subsequent streams. This process is best illustrated by the case of Snickers Gap, a prominent notch in the Blue Ridge of Virginia about fifteen miles south of Harpers Ferry. The floor of this notch hangs 700 feet above Shenandoah River at the west base of the ridge. A small stream called Beaverdam Creek heads near Snickersville at the east base, and flows eastward away from the ridge. The notch is a large one, and must have been cut by a large stream. Its reconstructed history is as follows: During an earlier cycle than the

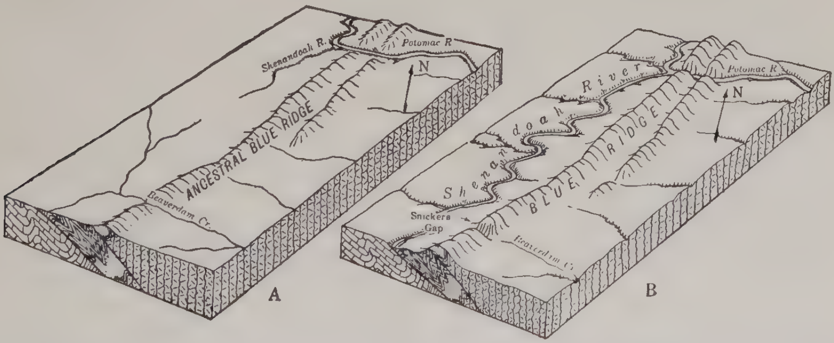


Fig. 273. — *A*, former drainage across the Blue Ridge in northern Virginia. *B*, present drainage resulting from the beheading of Beaverdam Creek by the Shenandoah.

one now in progress, when the land around the base of the Blue Ridge stood as high as the notch of Snickers Gap, this notch was occupied by a transverse stream, the ancestral Beaverdam Creek (Fig. 273, *A*). But the much larger transverse ancestral Potomac, which crossed the ridge 15 miles farther north, was able to cut its gap downward into the hard ridge-forming rocks much more rapidly. The Shenandoah, tributary to the Potomac through an easily eroded limestone area west of the ridge, kept pace with its master stream, and with the resulting favorable gradient, worked headward toward the south, cut through the divide which separated it from Beaverdam Creek, and " beheaded " the latter by diverting its upper waters toward the Potomac (Fig. 273, *B*). The gap was forthwith abandoned, and the beheaded and weakened creek found itself confined to the area east of the Blue Ridge. As the Shenandoah continued to cut downward, the ridge was etched out in greater relief and the abandoned gap was left high and dry.

Some wind gaps may have been formed by relatively weak streams which, affected by upwarps athwart their courses, were able to maintain their courses only for a time, having been later forced to abandon them and to flow along the strike.

The Evolution of Landscape

The varied landscapes about us attract our interest not only by their natural beauty but also by the clues they afford to a fascinating story of past events. Ability to decipher a part of this story by study of the land forms adds greatly to the enjoyment of scenery. Landscapes reflect the control of three principal factors: first, the character and structure of the bedrock; second, the climatic conditions under which the surface forms have developed; and third, the length of time that has elapsed since the development began. Different combinations of these three variable factors give rise to the seemingly endless variety of land forms.

All landscapes are ephemeral, and those of the present furnish an index to only the latest events in the history of the Earth. Other methods of exploring the ancient record are explained in the second part of this book.

APPENDIX A

MINERALS

INTRODUCTION AND DEFINITION

Minerals compose the crust of the Earth and are therefore among the most common objects of daily observation. A mineral may be defined as a naturally occurring substance that has a definite chemical composition and a definite combination of physical properties. This eliminates artificial products of the laboratory which may conform to the latter part of the definition. It also eliminates the natural products of organic agencies, as they do not show the definite chemical and physical characters of a mineral.

Minerals are composed of chemical elements. Some consist of single elements, such as diamond and graphite (different forms of carbon), or gold and silver so far as they occur in the free state in nature; but most minerals are made up of two or more elements united in such a way as to give a product that differs in its properties from any of the elements composing it.

At the present time about ninety different elements are recognized, but less than half of them are common and it has been calculated that more than 99 per cent of the crust of the Earth is composed of the following fourteen:

Oxygen, O	49.77	per cent
Silicon, Si	26.09	"
Aluminum, Al	7.34	"
Iron, Fe	4.11	"
Calcium, Ca	3.19	"
Magnesium, Mg	2.24	"
Sodium, Na	2.33	"
Potassium, K	2.28	"
Hydrogen, H	0.95	"
Titanium, Ti	0.39	"
Carbon, C	0.18	"
Chlorine, Cl	0.21	"
Phosphorus, P	0.10	"
Sulphur, S	0.10	"
All others	0.72	"
Total	100.00	"

Three elements together with their symbols, not included in the above list, are added here because they are present in certain of the minerals to be studied. Each one composes but a small fraction of one per cent of the Earth's crust.

Copper, Cu
Lead, Pb
Zinc, Zn

The number of minerals formed by the combination of even these few elements is very great but the common ones are relatively few.

CHARACTERS OF MINERALS

Chemical Composition

It was said above that minerals have a definite chemical composition. This composition, as determined by analysis, serves to define and distinguish the species, and indicates their relations to each other. Individual minerals react differently to various chemical reagents, and these reactions are one means of determining the kind of mineral under examination. It is beyond the scope of this discussion to treat that aspect of *mineralogy;* but there are many text books that treat the subject fully.

Physical Characters

Structure of Minerals. — Commonly the structure of minerals refers to their outward shape and form. The following descriptive terms are used in this connection, some of which are self-explanatory: *crystallized,* in definite crystals; *columnar; fibrous; botryoidal,* having a group of small rounded forms like a bunch of grapes; *reniform* (kidney-like) and *mammillary,* similar to botryoidal but in larger masses; *foliated* and *micaceous,* occurring in thin sheets; *granular,* in coarse to fine grains; *compact; earthy; stalactitic,* formed in stalactite masses similar to icicles; *massive,* showing compact material with an irregular form; *oölitic,* formed of small, rounded grains which resemble fish roe.

Crystals. — The majority of minerals under favorable conditions will form in crystals. These are bodies which are bounded by plane surfaces that are arranged according to definite laws of symmetry. The division of mineralogy known as crystallography is important and interesting but one whose detailed study takes considerable time. Certain principles will be pointed out here however. Perfect crystals are exceptional things. The great majority of mineral specimens will not show them. When they are to be observed, however, they will help materially in the identification of the mineral. In general the crystals of a certain mineral will show like or similar habits of crystallization. For instance, the mineral galena, PbS, characteristically crystallizes in cubes. Magnetite, Fe_3O_4, commonly occurs in eight-sided crystals that are called octahedrons. Garnets commonly occur in either dodecahedrons or trapezohedrons. These crystal habits are characteristic of these minerals and when recognized greatly aid in their identification.

Cleavage and Fracture. — The manner in which some minerals break is characteristic. If the break occurs with a smooth, plane surface the mineral is said to have a *cleavage*. This cleavage always takes place along planes, which may or may not be parallel to crystal faces. Some minerals will show but one cleavage, others two, three, or even six different cleavage planes. The number of planes of cleavage which a mineral shows and their relations to each other help to determine the mineral. Good examples are the cubic cleavage of galena (in three planes at right angles to each other), the rhombohedral cleavage of calcite (three planes not at right angles so that the resulting form is rhombic), the basal cleavage of mica (in one direction only). If a mineral does not show a cleavage it is said to have a fracture. Various kinds of fracture are as follows: *conchoidal* if the fractured surface is curved like the interior of a clam shell; *fibrous* or *splintery* if it shows a fibrous character; *uneven* or *irregular* if the surfaces are rough.

Color. — The color of a mineral is one of its most conspicuous physical properties. The color of many minerals is a definite and constant property and serves as an important means of identification. For example, the golden-yellow color of chalcopyrite, $CuFeS_2$, the blue-gray of galena, PbS, the black of magnetite, Fe_3O_4, are striking properties of these minerals. However, surface alterations may change the color of a mineral, as is shown in the golden tarnish frequently observed on pyrite, FeS_2. In noting the color of a mineral, therefore, a fresh surface should be examined. Moreover, many minerals show a variation in color in the different specimens. This may be due to a change in composition such as the gradual substitution of iron for zinc in the mineral sphalerite, ZnS, with the consequent darkening of the color of the mineral; or to impurities such as the red color given to quartz, SiO_2, by the admixture of hematite, Fe_2O_3. Other minerals such as fluorite, CaF_2, show a wide range in color without any apparent change in composition.

Color of Powder or Streak. — *The color of the streak* is an important aid to the identification of minerals. The streak is a thin layer of the powder of the mineral obtained by rubbing it upon an unglazed porcelain plate known as a streak plate. The color of the streak may be similar to the color of the mineral or quite different. For example, some varieties of hematite, Fe_2O_3, have a brilliant black color but give a red-brown streak that positively identifies the mineral.

Luster. — The luster of a mineral is the appearance of its surface due to the manner in which it reflects light. This must not be confused with color for two minerals with the same color may have totally different lusters just as a black paint with a shiny finish, such as an enamel, has an appearance different from that of a black paint with a dull finish because it reflects light differently.

Various descriptive terms are applied to the different kinds of luster exhibited by minerals. A partial list including the more important ones is given below:

Metallic. Having the appearance of a metal. Example, pyrite, FeS_2. (Most of the minerals with a black or dark-colored streak are included here.)

Vitreous. Having the luster of glass. Example, quartz, SiO_2.

Resinous. Having the appearance of resin. Example, sphalerite, ZnS.

Pearly. Having the iridescence of pearl. Example, some varieties of dolomite, $CaMg(CO_3)_2$.

Greasy. Looking as if covered with a thin layer of oil. Some varieties of massive silica, SiO_2.

Silky. Like silk, as the result of a fine fibrous structure. Example, fibrous gypsum, $CaSO_4.2H_2O$.

Adamantine. Having brilliant luster like that of a diamond, C.

Hardness of Minerals. — Minerals vary widely in their hardness, and a determination of this property is often an important aid to their identification. The relative hardness of any mineral may be told by comparing it with a series of minerals that has been chosen as a scale. The scale consists of crystallized varieties of the following minerals, each species being harder than those preceding it in the scale.

Scale of Hardness

1. Talc.
2. Gypsum.
3. Calcite.
4. Fluorite.
5. Apatite.
6. Orthoclase.
7. Quartz.
8. Topaz.
9. Corundum.
10. Diamond.

The relative hardness of any mineral in terms of this scale is determined by finding which ones of these minerals it can and which it cannot scratch. In making the determination the following precautions must be observed. Sometimes when a mineral is softer than another, portions of the first will leave a mark on the second which may be mistaken for a scratch. It can be rubbed off, however, whereas a true scratch will be permanent. Some minerals are commonly altered on the surface to material which is much softer than the original mineral. A fresh surface of the specimen to be tested should therefore be used. Sometimes the physical structure of a mineral may prevent a correct determination of its hardness. For instance, if a mineral is pulverulent, granular or splintery in its structure, it may be broken down and apparently scratched by a mineral much softer than itself. It is always advisable when making the hardness test to confirm it by reversing the procedure, that is, by rubbing the unknown on the material of known hardness.

The following materials will serve in addition to the above scale. The finger nail is a little over 2 in hardness, since it can scratch gypsum and not calcite. A copper coin is about 3 in hardness, since it can scratch calcite. The steel of an ordinary pocketknife is just over 5, and ordinary glass has a hardness of 5.5.

Specific Gravity. — The specific gravity of a substance is stated as a number that indicates how many times heavier a given volume of the material is than an equal volume of water. Minerals show a range of specific gravity from about 1.5 to 20. The great majority of minerals range between 2.0 and 4.0. There are various instruments that enable one to determine the specific gravity of a mineral with more or less accuracy, but for ordinary purposes it is sufficient simply to judge the weight of a fair sized piece in the hand. After some practice rather small differences in specific gravity can be detected in this way and a mineral approximately located in respect to this property.

COMMON MINERALS

Only a few of the more common minerals will be described. The student should always compare these descriptions with as many different specimens of the minerals as possible and should note the form, color, and luster of each sample and make the simple tests that determine the hardness, streak, and specific gravity.

MAGNETITE. An oxide of iron, a combination of ferrous and ferric oxides, $FeO.Fe_2O_3$, or Fe_3O_4.

Physical Characters. Color black. Streak black. Hardness 6. Heavy. Strongly magnetic. Usually granular or massive. Occurs in octahedral crystals. An important iron ore.

HEMATITE. The ferric oxide of iron, Fe_2O_3.

Physical Characters. Color reddish brown to black. Streak light to dark red-brown (Indian-red). Hardness 5.5–6.5. Commonly in botryoidal to reniform shapes with radiating structure. Often earthy. At times micaceous. Rarely in crystals. Abundant ore mineral of iron.

LIMONITE. Hydrous ferric oxide, $Fe_2O_3.H_2O$.

Physical Characters. Color dark-brown to nearly black. Streak yellowish brown. Hardness 5–5.5. Medium heavy. Commonly in mammillary to stalactitic forms with radiating fibrous structure; sometimes earthy. A minor source of iron.

PYRITE. Iron sulphide, FeS_2.

Physical Characters. Color pale brass-yellow. Streak black. Hardness 6–6.5 (unusually hard for a sulphide). Heavy. Usually granular. Sometimes in crystals, commonly striated cubes or octahedrons.

CHALCOPYRITE Copper-iron sulphide, $CuFeS_2$.

Physical Characters. Color golden-yellow; often tarnished to bronze or iridescent colors. Streak greenish black. Hardness 3.5 (note difference from pyrite). Heavy. Usually compact massive, rarely in crystals. A common and important ore mineral of copper.

SPHALERITE. Zinc sulphide, ZnS. Nearly always contains a small amount of iron.

Physical Characters. Commonly yellow brown to dark brown in color. Darkens with increase of iron. Resinous to submetallic luster. Hardness 3.5–4. Heavy. White to yellow and brown streak, always a lighter shade than the mineral itself. Perfect dodecahedral cleavage. Usually massive cleavable. The most important source of zinc.

GALENA. Lead sulphide, PbS.

Physical Characters. Color lead-gray. Streak grayish black. Hardness 2.5 (soft). Very heavy. Bright metallic luster. Perfect cleavage in three planes at right angles to each other, forming cubes. Often also in natural cubic crystals. The most important source of lead.

CALCITE. Calcium carbonate, $CaCO_3$.

Physical Characters. Color usually white or colorless. May be variously tinted, gray, red, green, blue, etc. Usually transparent to translucent. Hardness 3. Light in weight. Perfect cleavage in three planes at oblique angles to each other, giving rhombic-shaped faces (rhombohedral cleavage). Often in crystals which generally have a hexagonal cross section. Will effervesce freely on application of a drop of cold acid. This will serve to distinguish calcite from dolomite, $CaMg(CO_3)_2$, another common carbonate, which will not show effervescence under these conditions.

GYPSUM. Hydrous calcium sulphate, $CaSO_4.2H_2O$.

Physical Characters. Usually white or colorless. Hardness 2 (easily scratched with the finger nail). Light in weight. Has one very perfect cleavage. May be in tabular diamond-shaped crystals or in cleavable masses. Often also fine granular, sometimes fibrous.

HALITE (COMMON SALT). Sodium chloride, NaCl.

Physical Characters. Usually white or colorless. Hardness 2.5. Light in weight. Perfect cleavage in three planes at right angles to each other (cubic cleavage). Transparent to translucent. Salty taste. Generally in cubic crystals or in masses showing cubic cleavage.

QUARTZ. Silicon dioxide, SiO_2.

Physical Characters. Usually colorless or white, but frequently colored by different impurities, yellow, red, pink, amethyst, green, blue, brown, black. Vitreous luster. Transparent to opaque. Hardness 7. Light weight. Conchoidal fracture. Commonly in hexagonal crystals. The triangular faces at the ends of the crystals are usually smooth while the rectangular faces about the middle of the crystals are horizontally striated. Also massive.

Varieties. There are many varieties of quartz to which different names are given. A few are as follows: *Rock crystal*, colorless quartz, commonly in distinct crystals; *Amethyst*, quartz colored purple or violet; *Rose Quartz*, usually massive with a pink color; *Smoky Quartz*, quartz with a smoky yellow to brown or almost black color; *Chalcedony*, finely fibrous material, translucent with a waxy luster; *Agate*, a variegated chalcedony often delicately banded with different colors; *Jasper*, extremely fine grained quartz colored red with hematite.

GARNET. There are several different garnets which vary from each other in the elements they contain. They are all silicates with somewhat similar formulas. The most common garnet contains ferrous iron and aluminum, $Fe_3Al_2(SiO_4)_3$. Other garnets contain magnesium, calcium, manganese, ferric iron, chromium, etc.

Physical Characters. Color varies with the composition. Most commonly red or brown. May be yellow, white, green, black. Transparent to almost opaque. Hardness 7. Medium heavy. Usually distinctly crystallized, either in a form showing twelve rhombic-shaped faces or twenty-four trapezium-shaped faces.

ORTHOCLASE (A FELDSPAR). Potassium aluminum silicate, $KAlSi_3O_8$.

Physical Characters. Colorless, white, gray, flesh-red, more rarely green. Streak white. Hardness 6. Light in weight. Has two good cleavages making 90-degree angles with each other (whence name of mineral). Sometimes in crystals.

PLAGIOCLASE FELDSPARS. Sodium-calcium aluminum silicates.

Physical Characters. Various shades of gray, less commonly white. Transparent to opaque. Hardness 6. Light in weight. Have two cleavages making nearly a 90-degree angle with each other. Commonly distinguished from orthoclase by the color, by the presence on cleavage surfaces of a series of fine parallel striation lines, or by a bluish opalescence. Sometimes crystallize in thin bladed crystals with a curved surface and a pearly luster.

MUSCOVITE (COMMON LIGHT-COLORED MICA). A complex silicate containing potassium and aluminum.

Physical Characters. Possesses a perfect cleavage in one direction which allows the mineral to be split into excessively thin sheets. The folia are flexible and elastic. Transparent and almost colorless in thin sheets. In thicker blocks, opaque with light shades of brown and green. Hardness 2–2.5. Light in weight. Structure foliated in large to small sheets, sometimes in scales.

BIOTITE (COMMON DARK-COLORED MICA). A complex silicate containing potassium, magnesium, and aluminum.

Physical Characters. Perfect micaceous cleavage. Folia flexible and elastic. Color usually dark-green and brown to black. Thin sheets usually have a smoky color (differing from the almost colorless muscovite). Hardness 2.5–3. Light in weight.

CHLORITE. Complex silicate containing magnesium and aluminum. A complex group of minerals of similar characters which are called collectively the Chlorites from their common green color.

Physical Characters. Perfect micaceous cleavage. Folia flexible but not elastic (differing from muscovite and biotite). Color green of various shades. Hardness 2–2.5. Light in weight.

SERPENTINE. A magnesium silicate, $H_4Mg_3Si_2O_9$.

Physical Characters. Olive-green or yellow-green to blackish-green. Luster greasy or wax-like, silky when fibrous. Hardness 2.5–5, usually 4. Light weight. Usually massive but also fibrous or felted.

PYROXENE. A silicate containing chiefly calcium and magnesium; also varying amounts of aluminum, iron, sodium, etc.

Physical Characters. Color usually from light to dark green varying with amount of iron. Also at times nearly white or black. Transparent to opaque. Hardness 5–6. Light in weight. Often in prismatic crystals with eight sides. The angle between alternate faces is nearly 90 degrees (Fig. 274). These faces will fit into the corner of a box or tray. It is by these angles that the mineral can best be told from amphibole. Some specimens show a fair cleavage parallel to the faces lettered *m* in the figure, the angle between the cleavage faces being also nearly 90 degrees.

AMPHIBOLE. Silicate of calcium and magnesium with varying amounts of aluminum, iron, sodium, etc. Similar to pyroxene.

Physical Characters. Color usually light to dark green varying with amount of iron. Also nearly white or black. Transparent to opaque. Hardness 5–6. Light in weight. Often in prismatic crystals with six sides. Figure 275 shows that the angles between the faces lettered *m* are 124° and 56° (very different from the corresponding angles in pyroxene). Has a good cleavage parallel to the faces lettered *m*. The differences between the crystals and the cleavage angles in pyroxene and amphibole, and the fact that amphibole has the better cleavage constitute the chief distinctions between the two. Amphibole usually has a higher luster and a smoother surface than pyroxene. In some varieties of amphibole the crystals are long and needlelike, resulting in a fibrous structure. Pyroxene does not occur in this form.

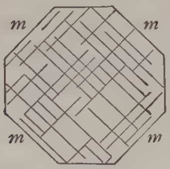


Fig. 274. — Cross-section of pyroxene normal to the long axis of the crystal. The cleavage traces are parallel to the prism faces (marked *m*). The alternate faces (those marked *m* or those unmarked) make angles of approximately 90° with each other. The cross-section has eight sides.

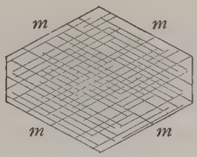


Fig. 275. — Cross-section of amphibole normal to the long axis of the crystal. The cleavage traces are parallel to the prism faces (*m*) which makes angles of 124° and 56° with each other. The cross-section has six sides.

Hornblende is a common dark variety of amphibole.

Olivine. Silicate of magnesium and iron, $(Mg,Fe)_2SiO_4$.

Physical Characters. Usually shades of green, more rarely brownish. Transparent to opaque. Hardness 6.5–7. Vitreous luster. Usually in granular, friable masses, more rarely in perfect crystals.

APPENDIX B

CHRONOLOGY OF EARTH HISTORY

	Periods		Revolutions (Times of Important Crustal Deformation)	Per cent of Total Time	Dominant Life
Cenozoic	Quaternary	Recent		4%	Age of man.
		Pleistocene	Cascadian Revolution		
	Tertiary	Pliocene			Age of warm blooded animals and flowering plants.
		Miocene			
		Oligocene			
		Eocene	Laramide Revolution		
Mesozoic		Cretaceous		8%	Age of reptiles and first flowering plants.
		Jurassic			
		Triassic			Age of reptiles and medieval plants.
Paleozoic			Appalachian Revolution	20%	
	Carboniferous	Permian			Age of earliest land animals and of the first forests.
		Pennsylvanian			
		Mississippian			
		Devonian			
		Silurian			Age of shelled invertebrates and of first fishes.
		Ordovician			
		Cambrian			
Proterozoic		Keweenawan	Some folding and intrusion of granite	68%	
		Huronian			Age of primitive invertebrates without shells.
Archeozoic			Widespread intrusion of granite		
		Timiskaming			Age of most primitive minute and soft tissued life.
			Folding and local intrusion of granite		
		Keewatin			

Total time represented — 1,000,000,000 ± years.

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